University of Nevada

Reno

HYDROGEOLOGIC STUDY OF GROUNDWATER-SURFACE WATER INTERACTIONS AT TOPAZ LAKE, NEVADA

A Thesis Submitted in Partial Fulfillment of the Requirements for the Degree of Master of Science in Hydrology

by

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ABSTRACT

A hydrologic study was conducted to investigate groundwater-lakewater interactions around Topaz Lake, Nevada, where an alluvial aquifer is hydraulically connected with the off-stream storage reservoir. Basic objectives of the study were: 1) the estimation of aquifer characteristics; 2) the estimation of recharge to the aquifer from the surrounding mountains; 3) the determination of the flux into the aquifer from the lake; and 4) the potential for contamination of the lake or groundwater from the other. Interpretation was done through the use of a two-dimensional groundwater flow computer model.

While the lake was increasing in storage, aquifer recharge was determined to be 62 percent from the mountains and 38 percent from the reservoir substantiating the potential for degradation of the aquifer by the reservoir. Groundwater seepage to the reservoir represented only one percent of its total volume, thus, degradation of the reservoir by groundwater is unlikely.

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ABSTRACT

A hydrologic study was conducted to investigate groundwater-lake water interactions of an aquifer and Topaz Lake in Nevada. Study objectives were the estimation of aquifer characteristics and recharge and the potential for contamination of the lake or groundwater from the other. Interpretation was done with a two-dimensional groundwater flow computer model.

While lake storage was increasing, 62 percent of aquifer recharge was determined to be from the mountains and 38 percent from the reservoir substantiating possible aquifer degradation by the reservoir. Only one percent of total reservoir volume was from groundwater seepage, thus, reservoir degradation by groundwater is unlikely.

INTRODUCTION

BACKGROUND

In 1921 Topaz Lake was created on the west side of Antelope Valley, Douglas County, Nevada, as an off-stream storage reservoir for irrigation purposes (Figure 1). It is maintained by the Walker River Irrigation District for use by farmers downstream in Smith and Mason Valleys during peak irrigation water demand seasons. With the creation of this body of water in the headwaters of the Walker River Basin, people began to look at the area as a place to build a summer or retirement home and use the lake for recreational purposes. The area of prime interest was the northwest corner of Topaz Lake from the shoreline and extending up to U.S. Route 395 (Figure 2).

With the advent of this increased interest, three subdivisions were registered with a total of approximately 230 lots available for residences, a trailer park, and three lodges involved in restaurant, bar, and the gaming industry. Throughout this area, the only source of domestic water is from individual wells penetrating the groundwater reservoir. There are presently no sewer systems, resulting in all of the sewage produced being disposed of through the use of septic tanks and leach fields.

To date about 25 percent of the available lots have been developed. With the expanded popularity of the area,

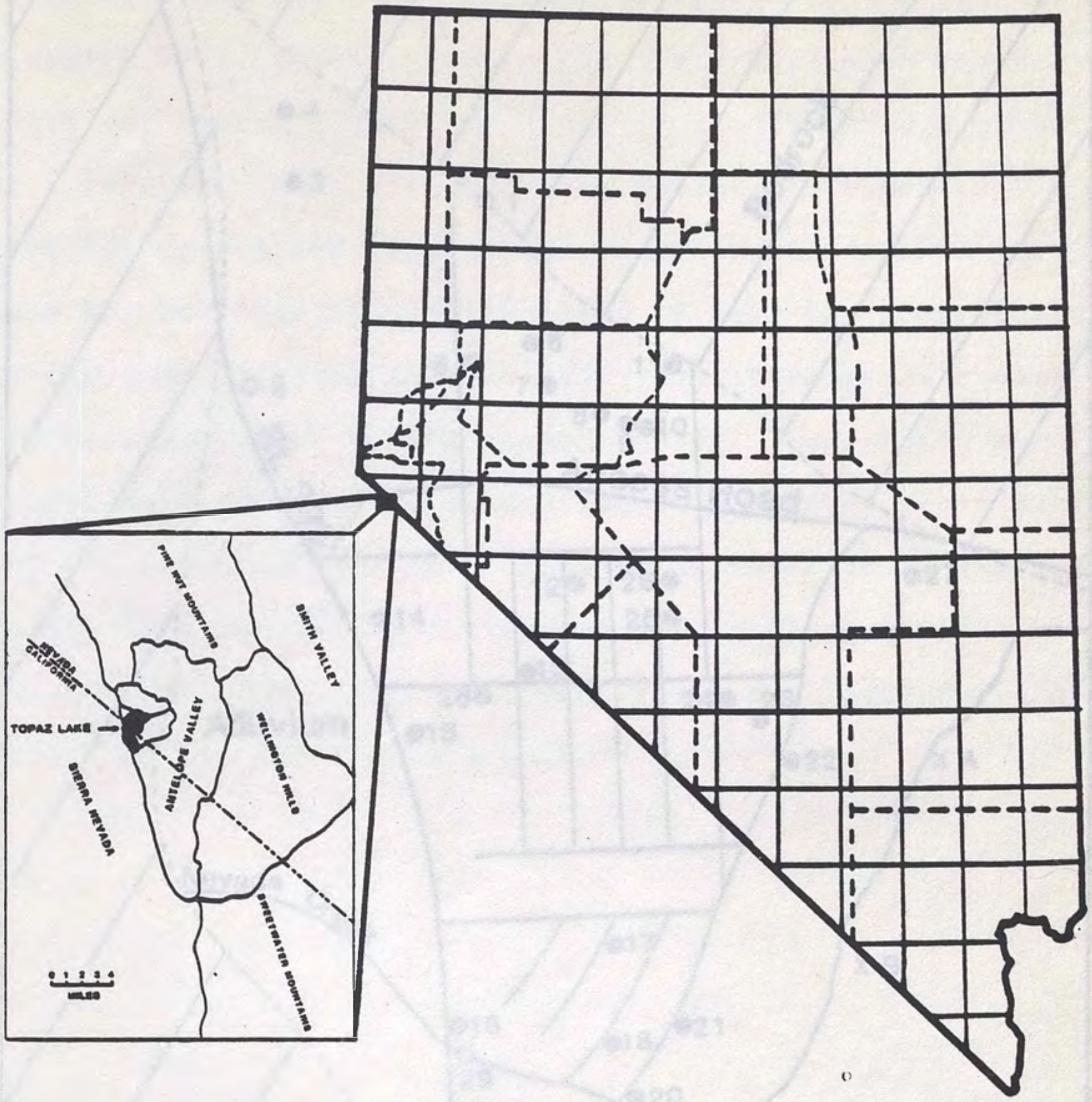

Figure 1. Location of Study Area.

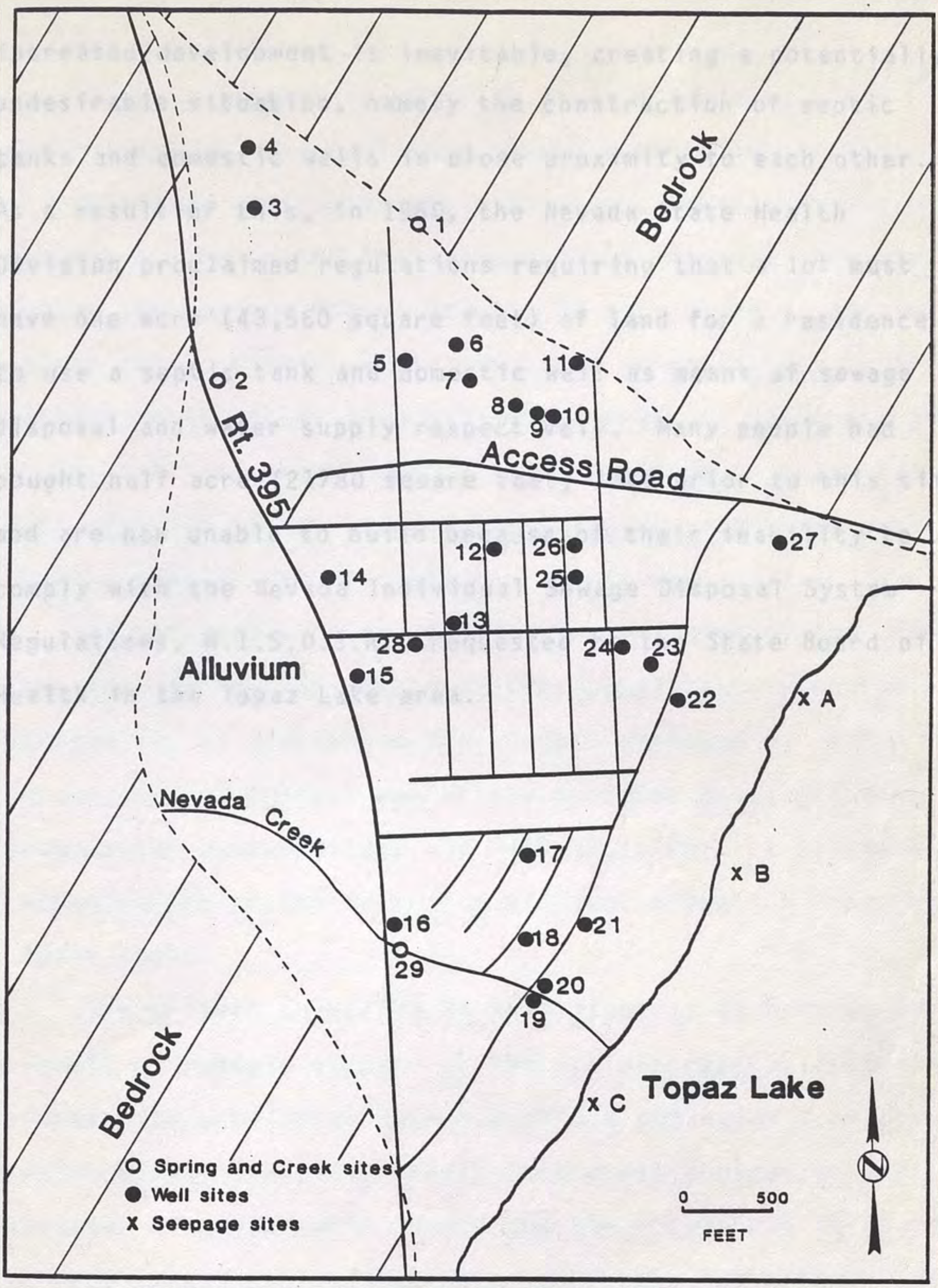

Figure 2. Study Area with Well, Spring, and Seepage Sites.

increased development is inevitable, creating a potentially undesirable situation, namely the construction of septic tanks and domestic wells in close proximity to each other. As a result of this, in 1960, the Nevada State Health Division proclaimed regulations requiring that a lot must have one acre (43,560 square feet) of land for a residence to use a septic tank and domestic well as means of sewage disposal and water supply respectively. Many people had bought half acre (21780 square feet) lots prior to this time and are now unable to build because of their inability to comply with the Nevada Individual Sewage Disposal System Regulations, N.I.S.D.S.R., requested by the State Board of Health in the Topaz Lake area.

PURPOSE AND SCOPE

With an increasing emphasis on maintaining a clean environment by many agencies, possible contributors to pollution are undergoing close scrutiny. Nearly twenty-nine percent of the United States population disposes of their wastes through the use of private on-site disposal systems, mainly septic tanks. These discharge approximately three billion cubic meters (800 billion gallons) of waste per year to the soil and are therefore a prime source of potential contamination of the groundwater below an area where septic tanks are widespread (Scalf, and others Aug, 1977).

At the present time, the northwest sector of the Topaz Lake area is not considered densely populated. With the eventuality of widespread development promoted by real estate agents, concern has arisen over the possibility of groundwater contamination and with this, the likelihood of contamination of the lake from effluent migration from the septic tanks.

The primary objective of this study is to provide an overall hydrologic picture of the storage reservoir-aquifer system with consideration for possible pollution from the septic tanks. Specific areas of interest include determination of groundwater levels and their response to change in lake stage, the evaluation of the phase lag between lake stage and well levels, estimation of aquifer characteristics, and the estimation of groundwater-reservoir interaction. This information will help in the regulation, planning and development of future land use around Topaz Lake.

The initial phase of this investigation involved the gathering of groundwater levels from selected domestic wells and existing United States Geological Survey (U.S.G.S.) observation wells during a one-year period. This was done monthly and was increased to weekly during seasonal reversals of lake stage. In addition to this, lake level stage data were compiled for concurrent days of observation of groundwater levels. This was necessary in order to determine the phase lag between a change in lake stage and the response of water levels in the wells upgradient from the lake. With this information, an estimation of aquifer characteristics in this area could be evaluated. After all of these data were collected, a computer program was used to reproduce the historic data collected.

Samples of groundwater and lake water were collected for chemical analyses from existing domestic wells located in the study area. Results of these were examined to determine trends in chemical composition and to see if any relationship existed between groundwater and lake water chemistry. These were also analyzed for contamination from septic tank effluent by checking for coliforms present in the samples.

LOCATION

The Topaz Lake region is in west central Nevada, at the western edge of the Basin and Range physiographic province. This area is characterized by north-trending mountain ranges and intermontane valleys filled with detrital material transported from the surrounding mountain ranges.

Topaz Lake is located along the west side of Antelope Valley whose center is approximately 36 miles south-southwest of the obtuse angle of the Nevada-California border with two-thirds in Douglas County, Nevada and one-third in Mono County, California (Figure 3).

The extension of Antelope Valley in which Topaz Lake lies is a topographically enclosed basin with an area of about 14 square miles (Rush and Hill, 1972). This basin has its lowest elevation of 4,913 feet at the bottom of the lake. It is encompassed by mountains on all but the southeast side. Here, at an altitude of about 5000 feet above sea level, there exists a narrow alluvial divide. In 1937, a low dam was constructed on the divide to further increase the storage of the reservoir.

On the west side of the lake, the Sierra Nevada locally rise to 9000 feet above sea level. Wild Oat Mountain forms the northern border and the Gray Hills border the reservoir on the east side. These mountains enclose the reservoir forming a basin into which the river flows.

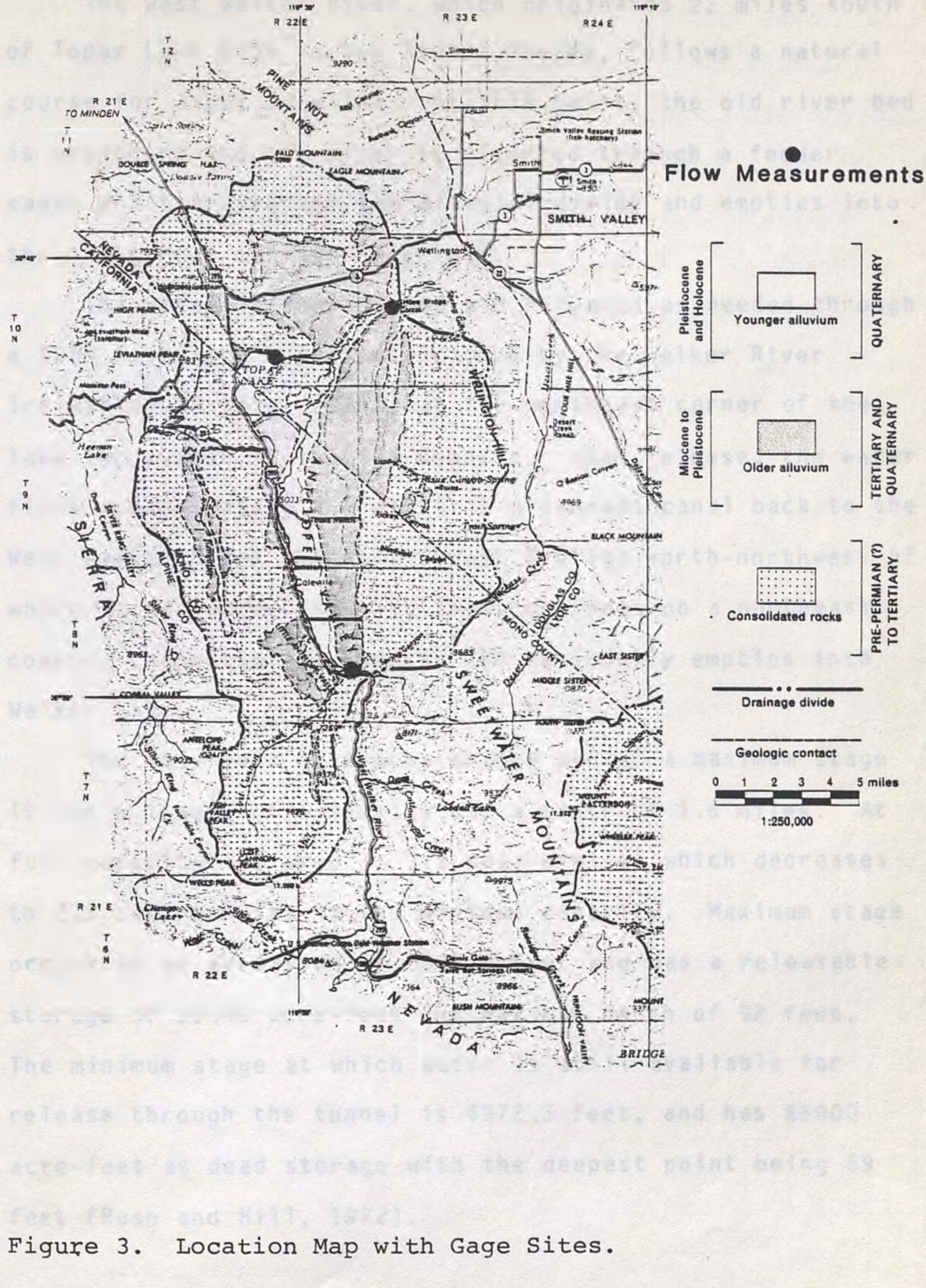

Figure 3. Location Map with Gage Sites.

The West Walker River, which originates 22 miles south of Topaz Lake high in the Sierra Nevada, follows a natural course for about 20 miles. At this point, the old river bed is abandoned and the river is diverted through a feeder canal which tranverses the alluvial divide and empties into the off-stream storage reservoir.

The water is then stored and released as needed through a 1200 foot long tunnel maintained by the Walker River Irrigation District (WRID) at the northeast corner of the lake for summer irrigation demands. Upon release, the water flows approximately 1.5 miles in a manmade canal back to the West Walker River channel, almost 5 miles north-northwest of where it was abandoned. It then continues on a northeast course, changes to southeast, and eventually empties into Walker Lake.

The reservoir is kidney shaped and at a maximum stage it has a length of 4.3 miles and a width of 1.6 miles. At full capacity its area is 3.8 square miles which decreases to 2.3 square miles at its minimum capacity. Maximum stage occurs at an evelation of 5005.0 feet and has a releasable storage of 59440 acre-feet and maximum depth of 92 feet. The minimum stage at which water is still available for release through the tunnel is 4972.3 feet, and has 65000 acre-feet as dead storage with the deepest point being 59 feet (Rush and Hill, 1972).

The project area is located on the northwest corner of the lake and covers an area of approximately 0.3 square miles. Within this area there are three lodges, one trailer park with 9 trailers, and approximately 46 houses. It is situated as a wedge bounded on the east by U.S. Route 395 and extends down to the Topaz Lake shoreline. The southern boundary is defined by the junction of the highway and the California-Nevada border. The gatehouse access road delineates the northern boundary (Figure 2).

CLIMATE AND VEGETATION

The Topaz Lake area is classified as an arid to semi-arid cold desert region. The Sierra Nevada and prevailing westerly winds are the major climatic controls. As moisture laden air moves west from the Pacific Ocean, it rises over the mountains and cools adiabatically. This results in condensation and heavy precipitation on the windward side of the Sierra Nevada, while the leeward slope receives little precipitation and is considerably drier. Consequently, this rain shadow effect causes a sharp contrast between the mean annual precipitation observed at Topaz Lake to the east, and in the mountains to the west.

The mean annual precipitation recorded at Topaz Lake during the years 1958-1968 and 1974-1979 was about 8 inches. To the southwest in the headwaters of the West Walker River, high in the Sierra Nevada, average annual precipitation can reach as high as 70 inches. Ths higher value occurs primarily during the winter months as rain or snow. The rest of the year experiences little precipitation usually in the form of localized convective thunderstorms caused by inequalities in temperatures at the boundary layer of the atmosphere. Distribution of precipitation from these storms is erratic, both in time and location.

Temperatures range from highs of 102°F to lows of -14°F. The mean annual temperature for the years 1974 to 1979 was about 50°F. The average January and July temperatures were 32°F and 69°F respectively.

The natural vegetation around the lake is typical of cold desert regions. On the valley floor around the lake, sagebrush (Artemesia sp.), bitterbrush (Purshia sp.), and rabbitbrush (Chrysothamnus sp.) thrive. Besides these shrubs, there are grasses and forbs found throughout the area. Along the high lake stage line, some cottonwoods are found. Pinyon pine and juniper stands cover the mountains to the north and west as elevations increase (Cooper, personal communication, 1980).

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PREVIOUS INVESTIGATIONS

Topaz Lake

The geology of the Topaz Lake area was mapped by Curtis (1951) as part of his Ph.D. thesis. Moore (1969) compiled a preliminary geologic map of Lyon, Douglas, and Ormsby Counties, Nevada, which included the study area. Most recent was a preliminary geologic map of the Walker Lake 1°x2° quadrangle by Carlson, et al. (1978).

The hydrology and geology of Antelope Valley were studied by Glancy (1971) as part of a water resources report. Groundwater data were compiled by Harrill and Nowlin (1976) for the northwest sector of Topaz as a basis for a future report. It encompassed a well inventory study and a preliminary well-level data collection. Nowlin (1976) gathered data pertaining to groundwater chemistry in the same area. Both of these reports were initiated to examine the possible effect of septic tank effluent on the groundwater and lake. Results of these studies provided background information for this question and indicated possible recharge from Nevada Creek and the lake to the groundwater system.

Rush and Hill (1972) compiled a bathymetric reconnaissance map of Topaz Lake. This also provided area and volume relationships as graphs for various values of lake stage.

Lake and Groundwater Interactions

In most studies of lake hydrology, the groundwater component of lake water is usually calculated as the residual in the budget equations. What is actually measured is the precipitation, evaporation, and surface water interactions with the lake which can lead to serious error and misinterpretation about lake and groundwater interactions. As a result of the relationship of groundwater to lakes being such a minor portion of hydrologic studies and therefore the least studied and least understood aspect of lake hydrology, some misconceptions of these interactions have come about. Besides this, inadequate data collection and instrumentation have resulted from a general lack of understanding of groundwater flow itself.

The various situations believed by investigators to exist around lakes are: (1) that either lakes are discharge points for adjacent groundwater systems and therefore lose water through their beds; or (2) that lakes are groundwater recharge points; or (3) that groundwater flow is a flow-through situation, in one side and out the other; or (4) that all three conditions exist.

Most studies of the interaction of groundwater and lake water had been done in areas of glacial terrane, the geologic environment in which most natural lakes occur. These studies were initiated to evaluate water budget data for one lake or multi-lake systems, or relationships between water table and prairie potholes, or sloughs of glacial origin.

Groundwater relationship to prairie potholes in North Dakota was studied by Sloan (1970, 1972) and Eisenlohr and others (1972). Observation wells were placed between potholes to determine the relationship between groundwater and pothole water levels.

Meyboom (1966) showed that small temporary sloughs in hummocky moraines are areas of discharge during summer and fall months but act as groundwater recharge sources during spring and early summer months. In another study by Meyboom (1967), permanent lakes in a hummocky moraine were found to be areas of groundwater discharge. This was determined through the use of a field technique (Langbein and others, 1951), which utilizes mass-transfer theory stating that the exchange of water vapor between a water surface and the atmosphere is directly proportional to the vertical humidity gradient and the wind speed. This mass-transfer analysis indicated that lake water was recharging the groundwater but upon looking at the groundwater flow system and its relationship to the lake, groundwater movement was found to be to the lake. What was occurring was that the lake was losing water to the phreatophyte zone surrounding the lake thus giving misleading results and not recharging the groundwater aquifer.

Several studies of the interaction between lakes and groundwater were carried out in Wisconsin by Hackbarth

(1968), Hennings (1974), and Possin (1973). The lakes studied in these reports were the flow through type where groundwater seeps into one side of the lake and the lake loses water to the groundwater on the other side.

In studies by Winter (1976, 1978), the general phsycial principles of the interaction of lakes and groundwater were analyzed. This analysis was done through the use of digital models simulating groundwater flow in vertical sections. Results of these simulations show that the movement of groundwater to and from a lake is dependent upon the continuity of a divide separating local from regional groundwater flow systems passing beneath the lake. Along this divide there exists a stagnation point where the head value is a minimum but is greater than the head represented by the lake level. In a situation where the divide is continuous, the lake cannot lose water to the groundwater system because the hydraulic gradient is towards the lake (Figure 4A). If the divide is not continuous, there exists an area where the gradient is away from the lake bed and the lake will lose water to the regional groundwater system (Figure 4B).

The study at Topaz Lake posed a unique situation for investigation. During part of the year while lake stage was declining, the study area was a groundwater discharge zone to the lake. Then when the lake stage began to rise from increased storage, the area became a groundwater recharge

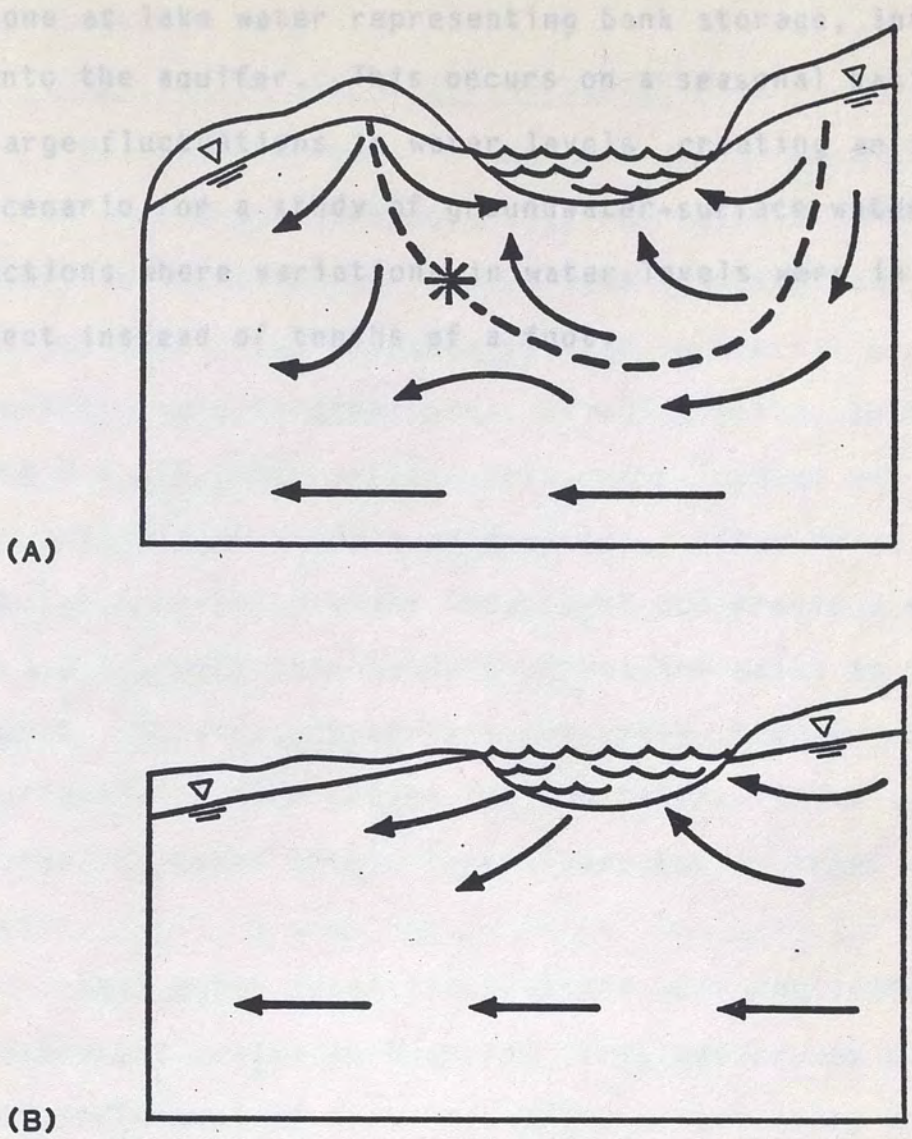


Figure 4. Groundwater Flow System,

A) With Stagnation Zone & Point,

B) Without Stagnation Zone & Point.

zone as lake water representing bank storage, infiltrated into the aquifer. This occurs on a seasonal basis with large fluctuations in water levels, creating an ideal scenario for a study of groundwater-surface water interactions where variations in water levels were in tens of feet instead of tenths of a foot.

METHODOLOGY

WELL MONITORING AND NUMBERING SCHEME

Wells were chosen in the area based on several criteria. In order to have a good representation of the area, location was a factor. Wells were selected that were dispersed throughout the area so that an overall picture of the aquifer could be developed. Existing wells, both domestic and U.S.G.S. observation wells, were located and those which were easily accessible were chosen. After doing this aerial photos from the Highway Department and previous work by the U.S.G.S., were then used to select the wells to be monitored. As a result of this, 26 wells, two springs, and one surface site were chosen for the study. These were monitored for water level fluctuations for a period of one year.

Well water level fluctuations were monitored using an instrument called an M-scope. This device was made with a 300 foot spool of two lead cable, marked every five feet with a band and attached to an electrode probe on the down-hole end. An ammeter on the spool indicated contact between the electrode and the water surface by measuring the current passed through the cable from four C batteries. This provided positive indication of contact with the water surface and therefore water level depth below land surface.

Monthly measurements of the water levels in the wells were taken while the lake stage was steadily increasing or decreasing. Once the outlet gate was closed in October to start winter storage, water level measurements in the wells were increased to a weekly basis. The increased measurements were necessary because as the lake level begins to rise, the water table stops declining, stagnates for awhile, then reverses and begins to rise.

When the lake level begins to decline due to an increase in demand downstream for irrigation water, the water table stops rising, and begins to decline, mimicing the lake stage pattern. This increased frequency of measurements during periods of peaks and troughs in water levels enables accurate determination of the phase lag which is necessary to evaluate aquifer properties.

Since most of the wells were domestic ones, there was the possibility that the readings were affected by recent pumping which would give erroneous values of water levels. To decrease the chances of this occurrence, water level measurements were taken at approximately the same day and time each week and during a period when there was less demand for water.

Well #28 was chosen for continuous monitoring of its water level fluctuations. A Stevens F-type water level recorder was placed within a fifty gallon drum which had been cut in half from top to bottom and mounted sideways on legs. The top was hinged and was locked to prevent theft or tampering. On the bottom of the drum a sleeve extending down to the ground was slipped over the well casing through which the weight and float were suspended.

The recorder was set up with selected gears and drum so that one revolution would equal a ten foot change in water level, and would cover one month before chart changing was necessitated. This unit was left in place from October 1979 to October 1980.

The collected water levels were converted to elevations above mean sea level, which was chosen as the common datum level. Several previous reports had varying data on elevations among them with an accuracy of plus or minus ten feet. Because of this variance and the necessity for accurate water levels so that comparisons could be made, it was decided to survey the elevations in with a surveying level.

Datum points for the survey were elevation markers surveyed in by the Highway Department during construction on Highway 395. These markers were located along the highway above the study area and were marked by bronze pins with stamped station numbers in a brass cap, all driven into the earth with a cement collar around it. From these markers, all of the wells used in this study were surveyed in during January 1980. All of the well elevations used in this report are accurate to plus or minus one foot.

The number assigned to a well or spring in this report acts as both an identification and location number. It is based on the index of hydrographic areas by Rush (1968) and on the Mount Diablo base and meridian of the General Land Office. A sample number consists of six units separated by spaces. The first unit is an assigned number ranging from one to twenty nine. The second unit is a number representing the hydrographic area in which it is located. The third and forth units represent the township and range respectively. The section number is found next as the fifth unit. This is then followed by a letter representing quarter sections. These sections can then be divided into 40, 10, or 2.5 acre tracts. If a number follows the letters, it shows the sequence in which the well was recorded in that tract.

For example, well 3 106 N10 E22 32 ABCD2 is on the location map as number 3. It is found in the hydrographic index area 106, called Antelope Valley, Nevada. This well was the second one recorded in the SE 1/4 of the SW 1/4 of the NW 1/4 of the NE 1/4 of section 32. It is also in Township 10 North and Range 22 East, Mount Diablo base line and meridian.

WATER CHEMISTRY DATA

Chemistry of the groundwater was studied to determine the relationships between the groundwater system and reservoir stage. The wells closer to the shoreline were given more attention since there was the possibility of seasonal gradient reversals and possible groundwater- lake-water exchanges.

Well water samples for gross analysis were collected initially at each well being monitored that was a pumping well. Water was allowed to run at least three minutes before sampling occurred. A one gallon cubitainer was filled and field temperature taken. This sample was then brought back to the Water Resources Center, Desert Research Institute's analytical laboratory where wet chemistry analyses were done. References and detection limits for these methods are listed in Table 1.

After the initial sampling in August and October, 1979, samples were again collected and analyzed in July 1980 at three sites close to the shoreline and one from the lake. This was done to see if there was any relationship between lake and groundwater chemistry which might indicate the existence of interactions between the two.

Bacteria samples were also collected from wells to see if they contained any coliforms. These were collected in 125 milliliter glass bottles treated with enough sodium thiosulfate to neutralize 15 parts per million chlorine.

Table 1.--Water Chemistry Analyses Performed by the WRC (Analysis Methods, References, and Detection Limits)

Constituent	Method	Reference	Detection Limits (mg/l)
pH	Electrometric	1	
TDS	Arithmetric summation of ionic constituents		
Specific Conductivity	Meter, Temperature-compensating	1	
Sodium	Direct Atomic Absorption	1	.2
Potassium	Direct Atomic Absorption	1	.1
Calcium	Direct Atomic Absorption	1	.2
Magnesium	Direct Atomic Absorption	1	.1
Bicarbonate	Electrometric Titration	1	1.0
Chloride	Ferricyanide-auto	1	.5
Sulfate	Turbidimetric	1	1.0
Nitrate	Brucinesulfate	1	.1
Silica	Molydateblue-auto	1	1.0

1. U.S. Environmental Protection Agency, 1979.

Table 2.--Bacteriological Results for Topaz Lake Well and Surface Samples

Sample ID	Map # on Fig. #2	Total Coliform[1] Colonies/100ml Sample	Fecal Coliform[1] Colonies/100ml Sample
TLB 1	26	0	0
TLB 2	25	0	0
TLB 3	12	0	0
TLB 4	13	0	0
TLB 5	15	0	0
TLB 6	16	0	0
TLB 7	17	0	0
TLB 8	20	0	0
TLB 9	21	0	0
TLB 10	23	0	0
TLB 11	8	0	0
TLB 12	9	0	0
TLB 13	D	4 (<10)	1
TLB 14	6	0	0
TLB 15	E	16 (<10)	7

1. Analyzed by membrane filtration. Values are the same from both the Bureau of Laboratories and Research, Nevada Division of Health, and the Bioresources Center, Desert Research Institute, Reno, Nevada, unless in brackets which were reported by the Nevada Division of Health.

RIVER FLOW AND LAKE STAGE DATA ACQUISITION

Flow data for Walker River were obtained from water-stage recording flow gages maintained by the U.S.G.S.. There were two gages on the river, one above the lake near Coleville, California and one below at Hoye Bridge near Wellington, Nevada. These data from the U.S.G.S. on flows is provisional data as it was received before final publication in the Water Resource Data for Nevada for this year.

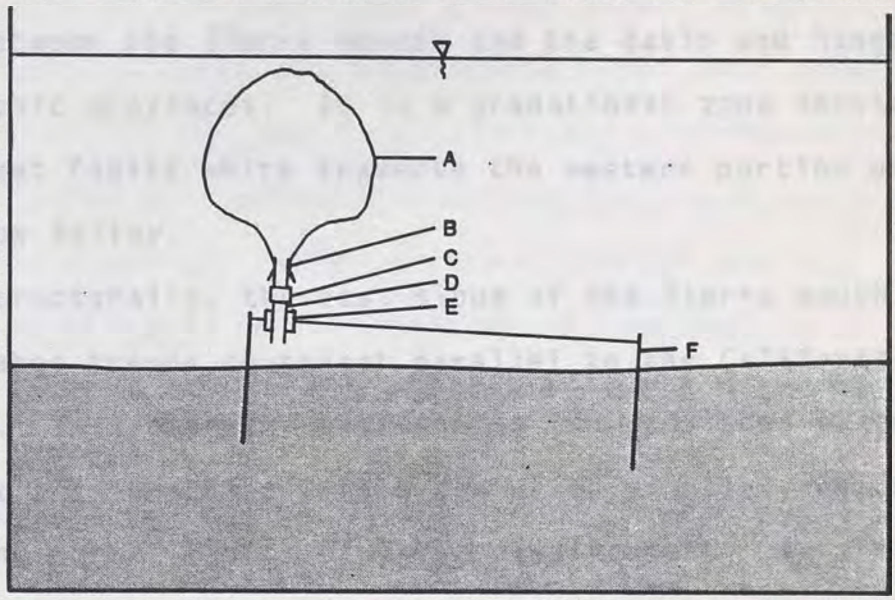
Lake stage and volume were available from float and nonrecording gages, read daily. The gage was located on the lake and is maintained by the Walker River Irrigation District (W.R.I.D.), which furnishes the elevations to the U.S.G.S.. See Figure 3 for location of gages.

SEEPAGE MEASUREMENTS

A device described by Lee (1977) was used to measure seepage flux and to collect water samples from infiltrating groundwater (Figure 5). It was constructed out of a 55 gallon drum opened completely on one end and sealed on the other. On the top, a hole with a one-way valve was made to which a deflated plastic bag was attached. The cylinder was about 12 inches high and was set open-end down, into the sediment at the bottom of the lake, until it was buried halfway.

To measure flow and collect a sample, the plastic bag was left on for about 3 hours on four separate days in September 1980, at one week intervals. This sampling was done while the lake and water table levels were declining and had been doing so for about two months. At this time the groundwater hydraulic gradient is towards the lake, draining the aquifer of water which had been bank storage from the lake.

Three sites were chosen at locations which were easily accessible and not too deep. The main problem was that with the lake level declining, they had to be placed in water deep enough so they would be both accessible and usable for a period of one month. Another problem was the rocky lake bottom which made it difficult to find a location where the cylinder could be emplaced 6 inches into the bottom. Figure 2 shows the approximate locations of the three cylinders used.



A- 3-liter, .75mil plastic bag

B- rubber-band wrap

C- one-way flow valve

D- .64cm ID polyethylene tube

E- one-hole rubber stopper

F- 15 X 57cm diameter cylinder
(end-section of a steel drum)

Figure 5. Seepage Meter Diagram.

GEOLOGY

GENERAL GEOLOGY

Topaz Lake is situated on the west side of Antelope Valley, Nevada, adjacent to the Sierra Nevada. This portion of the West Walker River basin is found in the transitional zone between the Sierra Nevada and the Basin and Range geomorphic provinces. It is a gradational zone denoted by the great faults which traverse the western portion of Antelope Valley.

Structurally, the east slope of the Sierra south of Lake Tahoe trends southeast parallel to the Califoria-Nevada border. This boundary between the two provinces becomes complex and irregular with a series of parallel, northerly-trending normal faults of large displacement. As a result, these faults formed similarly trending ranges as master fault-block ranges. The Virginia Range-Pine Nut Mountains-Wellington Hills- Sweetwater Range, represents one of these fault-block ranges bordering Antelope Valley on the east (Figure 1).

The intermontane valleys of the area are located between or along the faults where streams have been able to erode the weathered bedrock. A series of interbedded floodplain and stream channel deposits, alluvial fans, and playa sediments comprise the deposits of the groundwater basins. Many of these basins have high water tables as a result of enclosure of the valley by essentially impermeable materials and a gentle slope to the valley floor.

In the mountains surrounding Topaz Lake, the rocks exposed are mostly Mesozoic in age. These are found in the Gray Hills to the east, Wild Oat Mountain to the north, and to the west on the east flank of the Sierra Nevada. These early Mesozoic rocks are metamorphic which have been strongly eroded and now form relatively limited outcrops as roof pendants and septa. The surrounding rocks are part of the massive younger Sierra Nevada batholith and are Cretacous granitic rocks comprised of nonporphyritic quartz monzonites, granodiorites, and hybrid mafic rocks (Moore, 1969).

Through the use of fossils, the limited outcrops of Mesozoic metamorphics have been dated to be Late Triassic and Early Jurassic in age (Gilbert and Reynolds, 1973). Due to the deformation by, and heat from the granitic pluton, the pre-granitic sequence of rocks was slightly metamorphosed. Of the rocks involved in this deformation, the metavolcanics are more abundant than the metasedimentary rocks, although both are interbedded. The distinction between the two types of rocks on the geologic map was done solely on the basis of the prevalent lithologic unit in the body.

The Gray Hills and Wild Oat Mountain are both metavolcanic in origin and of Triassic-Jurassic age. These rocks originated from volcanic rocks which were metamorphosed into meta-andesites and metadacite, largely as volcanic breccia. On the west side of Topaz, the mountains are Jurassic-Triassic metasedimentary rocks consisting of shale, slate, tuffaceous siltstone, sandstone, graywacke, and some volcanic material.

LOCAL GEOLOGY

The area of study is located on a Quaternary alluvial fan on the northwest shoreline of Topaz Lake (Moore, 1969). It is bounded by consolidated metasedimentary rocks on the west and matavolcanics on the north (Moore, 1969, pl. 1). A north trending Quaternary normal fault separates this alluvium from the metasedimentary rocks to the west (Gilbert and Reynolds, 1973).

The alluvium was eroded and transported from the peripheral mountains and deposited as an alluvial fan. It is primarily composed of alluvial fan gravel, stream-laid gravel, sand, silt, and some talus material.

As this material was carried downslope, the larger, heavier constituents were deposited first on the steeper upper reaches of the fan. As the slope decreased the smaller detrital material was deposited. This action formed a sequence of clays, sands, gravel, and boulders throughout the area.

Available driller's well logs for the area were obtained from the office of the Nevada State Engineer. The majority of them are rather general in their description of materials encountered. In most of the logs, a series of clay and broken rock layers interspersed with gravel and sand lenses are found. The gravel and sand lenses occur as intermittent stringers of buried stream channel deposits.

PHYSICAL PROPERTIES AND WATER-BEARING CHARACTERISTICS OF THE ROCKS

Bedrock

The bedrock in the area is comprised of the metasedimentary and metavolcanic units described earlier. This material is generally untapped by the wells in the study area and therefore information on it is rather sketchy. The available well logs are very general in their description of lithological units encountered, making it difficult to discern if a well had penetrated the bedrock.

Well #4 is the deepest in the area, at a depth of 350 feet below land surface. At about 270 feet down, a hard brown rock layer 80 feet thick was penetrated. This material yielded little if any water and the well casing was not perforated below 250 feet. Of the well logs collected, this is the only one that penetrates the bedrock and its ability to provide water is rather limited. The only evidence of transmissivity in this rock is from two springs in the area, numbers 1 and 2 on Figure 2. Their flows were estimated at 1-2 gallons per minute (Harrill and Nowlin, 1976).

In some areas of intense structural deformation, some limited water might be yielded to wells from localized fracture zones. The chances of intercepting such zones are so poor that the bedrock, whether it be in the mountains or buried beneath the alluvium, is not considered to be a practical source of groundwater, and is, therefore, not discussed in this report.

Alluvial Material

The alluvial material is the major source of groundwater in the area. It is characterized by unconsolidated lenses of sand, gravel, clay, and silts in fluvial, stream channel deposits, and represents detrital material derived from the surrounding mountains.

In order to reach the groundwater, wells had to penetrate the alluvium to varying depths, depending on how far upslope from the lake they were. Well #21 is at an elevation of 5026 feet and penetrates only 80 feet of alluvium, with an average static water level that was 40 feet below land surface. On the other hand, well #16, at an elevation of 5127 feet, penetrates 216 feet of alluvium without encountering bedrock, and had an average static water level that was 138 feet below land surface.

The depth to groundwater in the wells is variable depending upon the lake stage. During the period of data collection, from July 1979 to October 1980, the lake was observed to vary a total of 28.5 feet between its high and low stages of the year. In response to this change, the groundwater levels also were observed to change about the same amount. The depth to water in well #21 was 52.5 feet in November 1979 but was only 26 feet below land surface in July 1980. This was a net change of 26.5 feet in water table elevation in response to a 28 foot change in lake stage.

Upon examining the plots of lake stage versus well level for wells in the area (Appendix I) a direct communication between the lake water and the groundwater system is evident. In well #21, located along the shoreline, a change in lake stage affects the water level in the well almost immediately as seen in Figure 6. In contrast to this immediate response, well #12, shown in Figure 7, shows that it does not respond to lake fluctuations for some time. This delay, or phase lag, in reaction time of the well in response to a change in lake stage, varies as a function of distance from the lake. In well #21, the phase lag is only a few days, whereas well #12 shows a phase lag of about twenty-seven days.

Evaluation of this phase lag throughout the area can be used to estimate the hydraulic conductivity of the aquifer. Using the plots in Appendix I of lake stage versus static water levels in the wells, an estimate of the phase lag between the time a lake pulse occurs and when a given well exhibits that change can be determined. Knowing this and the distance to the shoreline, a rate of travel through the aquifer, as length per unit time, can be obtained. An average estimate of the hydraulic conductivity, determined through the use of the phase lag, was found to be 60 feet per day. This value is a reasonable estimate for the hydraulic conductivity of relatively clean fan deposits.

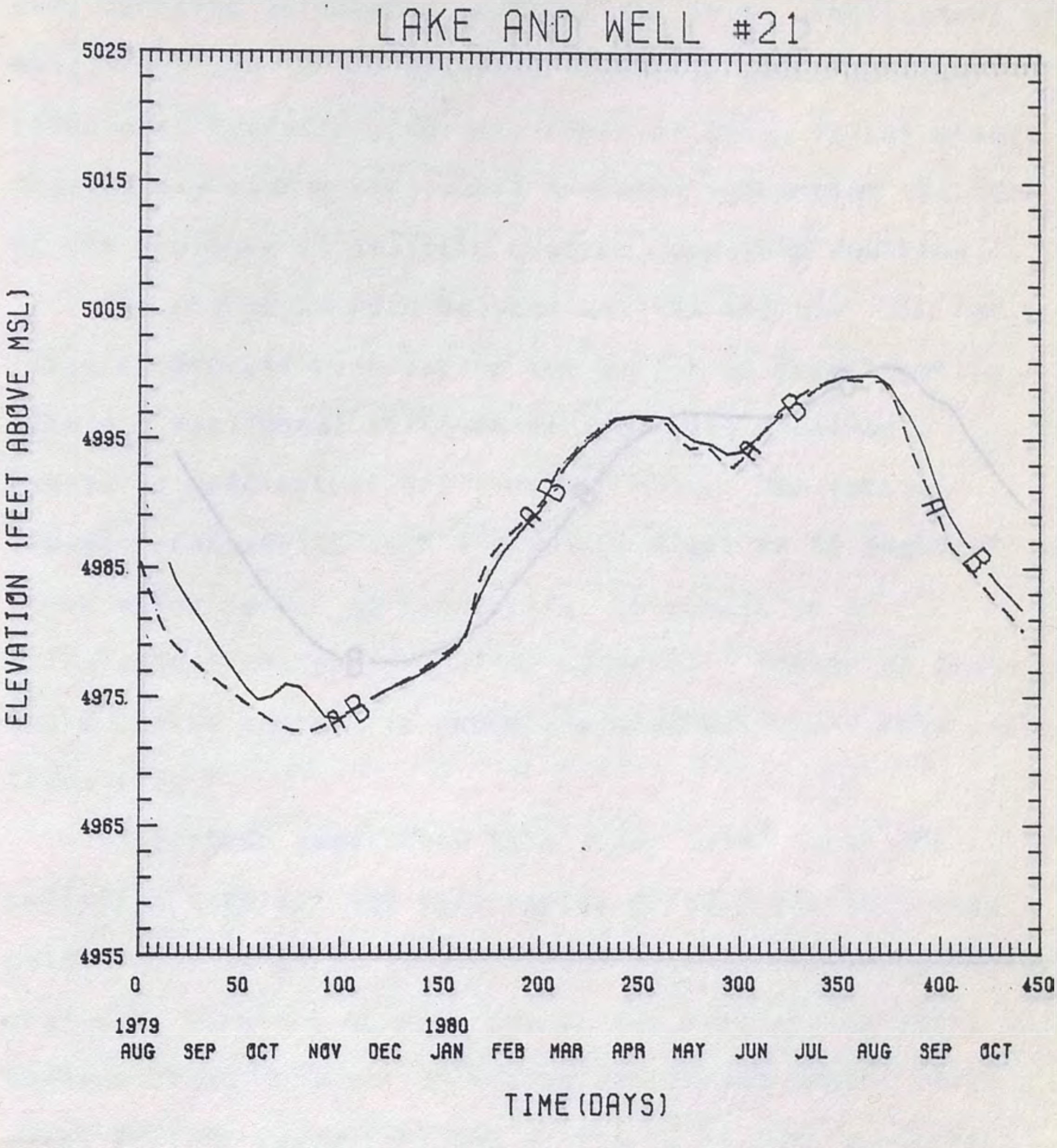


Figure 6. Superposition of Water Level in Well #21 and Lake Stage.

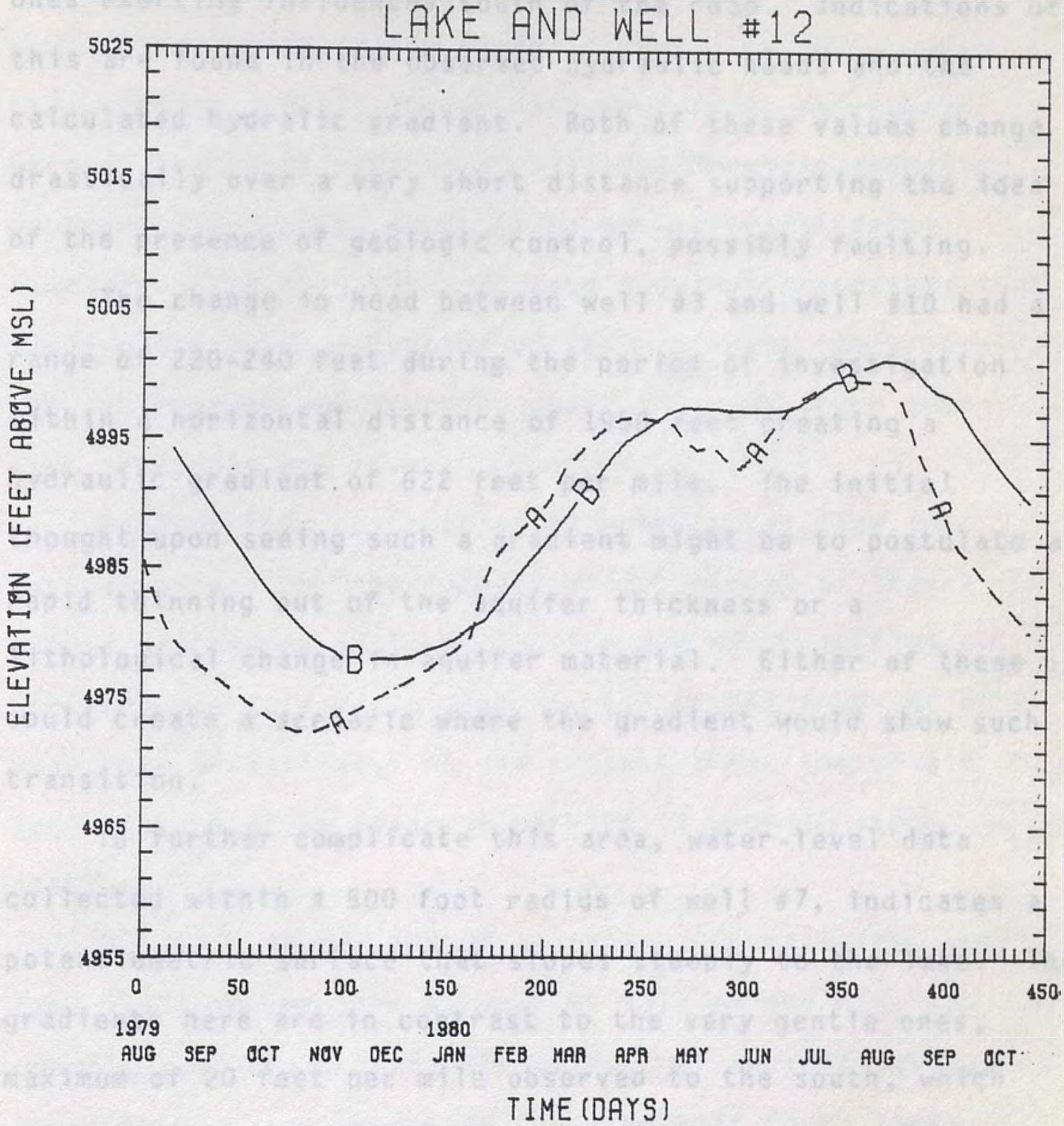


Figure 7. Superposition of Water Level in Well #12 and Lake Stage.

BARRIER ZONE TO GROUNDWATER FLOW

Wells located north of the east-west access road show evidence of having hydrogeologic controls in addition to the ones exerting influences south of the road. Indications of this are found in the observed hydraulic heads and the calculated hydralic gradient. Both of these values change drastically over a very short distance supporting the idea of the presence of geologic control, possibly faulting.

The change in head between well #3 and well #10 had a range of 220-240 feet during the period of investigation within a horizontal distance of 1950 feet creating a hydraulic gradient of 622 feet per mile. The initial thought upon seeing such a gradient might be to postulate a rapid thinning out of the aquifer thickness or a lithological change in aquifer material. Either of these could create a scenario where the gradient would show such a transition.

To further complicate this area, water-level data collected within a 500 foot radius of well #7, indicates a potentiometric surface that slopes steeply to the lake. The gradients here are in contrast to the very gentle ones, maximum of 20 feet per mile observed to the south, which gives further support to the existence of some subsurface geologic structure. One set of observations gave a gradient of 2745 feet per mile between well #6 and #7 which are only 250 feet apart but have hydraulic head change of 130 feet.

This high gradient could be interpreted as resulting from a band of rock having a very low hydraulic conductivity in the lateral direction or an area of faulting. Evidence supporting the idea of faulting is that the western boundary of the alluvium is delineated by a north-south fault (Moore, 1969), and an east-west fault was mapped through the lower portion of Wild Oat Mountain trending perpendicular to the north-south one at the alluvial boundary. The existence of these two faults suggests that there might be an extension of the east-west fault towards the north-south one, running under the area where these high gradients are found.

Without this obstruction to groundwater flow in the lateral direction, the damming effect needed to induce such a head difference would be lacking and the resulting potentiometric surface would slope gently to the south at an even rate of decline. At the present time, the origin and exact nature of an apparent barrier is open to speculation.

Well #5, #6, #7 and #8 all show influence from the lake fluctuations, not as much as well #9 and #10, but superimposed on this are the effects of whatever geohydrologic factors are producing the gradients found here. Because of the high gradient present, its unknown origin and lack of detailed structural and geologic information, the area north of the east-west access road will not be considered in the model.

Several wells north of the access road exhibited influence from storm recharge within 40 days of the occurrence of a particular storm on day #165. The change was the greatest at well #8 with the water level rising 28 feet, and the smallest at well #10 with a rise of 14 feet, of which both occurred within 7 days. Of all the wells monitored, wells #7, #8, #9, and #10 were the only ones showing a discernable influence from recharge by the storm. In going from well #8 to #10, downgradient towards the lake, the magnitude of the influence diminished. This was probably a result of a thickening of the alluvium and being farther away from the barrier zone mentioned earlier.

SURFACE WATER HYDROLOGY

WEST WALKER RIVER

The West Walker River supplies the majority of the water to Topaz Lake. Its head is in the Sierra Nevada in Mono County, California, approximately 22 miles south of Topaz Lake. A gaging station maintained by the U.S.G.S., 5 miles southeast of Coleville, California, in the NW 1/4 NE 1/4 section 28, T.8N., R.23E., Mono County, is used to measure flow past this point. The average annual discharge from a drainage area of 271 square miles, based on 27 years of record, is 179,000 acre feet.

This flow is mainly derived from snowmelt runoff in the higher elevations of the Sierra. Since snowmelt is the dominant factor with respect to runoff, an unregulated streamflow pattern is found above Topaz Lake (Figure 8). Peak flows can be expected in June or July, with lesser peaks of shorter duration caused by rainfall from November to March. Because of the inability of the steep, rocky eastern slope of the Sierra Nevada to retain rain or water from snowmelt, the fluctuations in observed streamflows at Coleville, California, can be extreme (See Figure 8).

After passing through Coleville, the river trends northward towards Topaz Lake. About 2 miles south of Topaz, California, a diversion in the river channel diverts streamflow from the natural channel through a canal into the reservoir. This inflow is stored in the reservoir and is

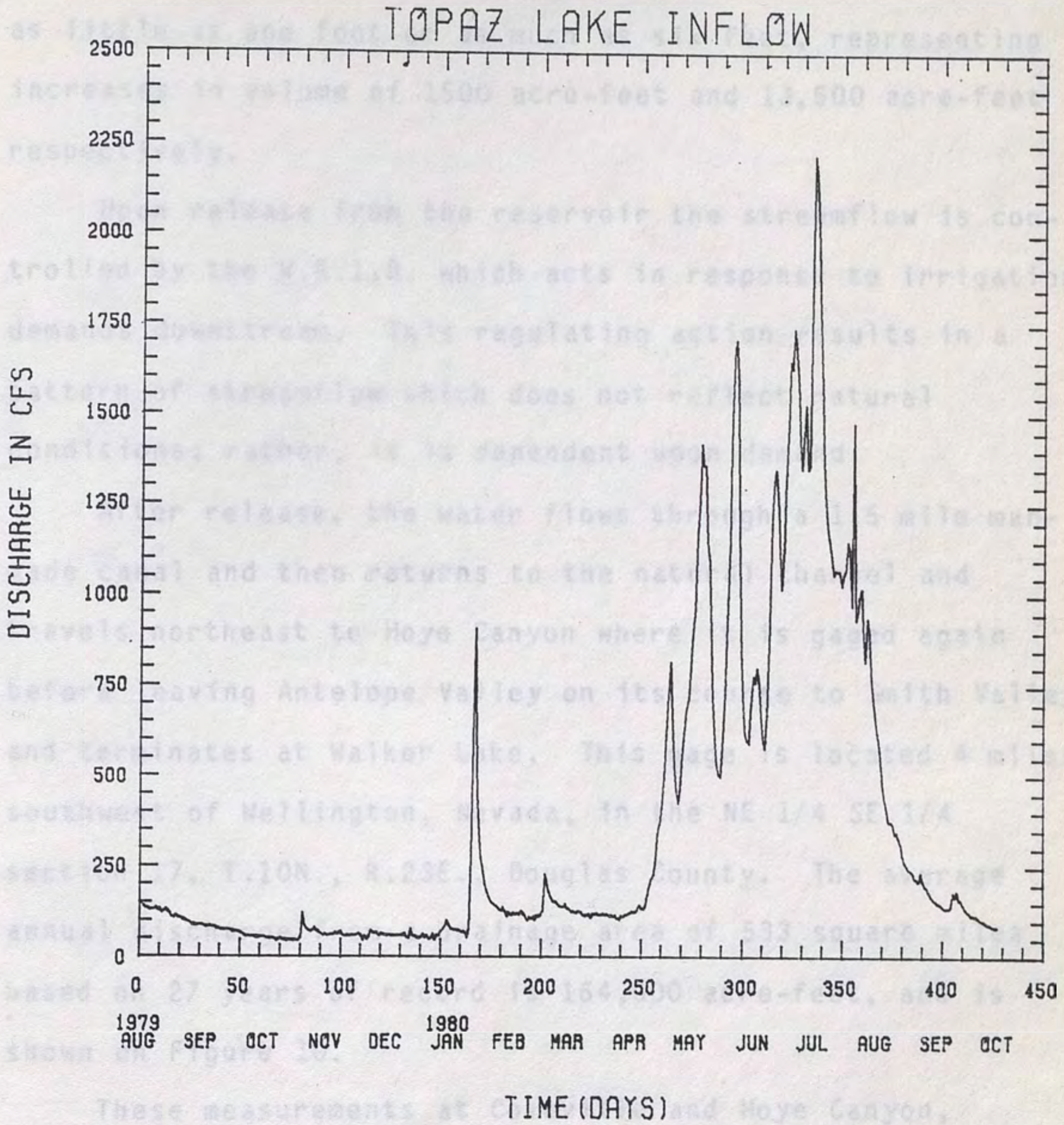


Figure 8. Topaz Lake Inflow.

released as needed downstream.

The unregulated inflow causes the lake stage to be capable of rising rapidly over a short period of time. Figure 9 shows that in one month the lake stage could change as little as one foot or as much as six feet, representing increases in volume of 1500 acre-feet and 13,500 acre-feet respectively.

Upon release from the reservoir the streamflow is controlled by the W.R.I.D. which acts in response to irrigation demands downstream. This regulating action results in a pattern of streamflow which does not reflect natural conditions; rather, it is dependent upon demand.

After release, the water flows through a 1.5 mile man-made canal and then returns to the natural channel and travels northeast to Hoye Canyon where it is gaged again before leaving Antelope Valley on its course to Smith Valley and terminates at Walker Lake. This gage is located 4 miles southwest of Wellington, Nevada, in the NE 1/4 SE 1/4 section 17, T.10N., R.23E., Douglas County. The average annual discharge from a drainage area of 533 square miles based on 27 years of record is 164,000 acre-feet, and is shown on Figure 10.

These measurements at Coleville and Hoye Canyon, represent inflow and outflow to Antelope Valley respectively. Table 3 shows the annual flow fluctuations of inflow and outflow to this area for the water year periods

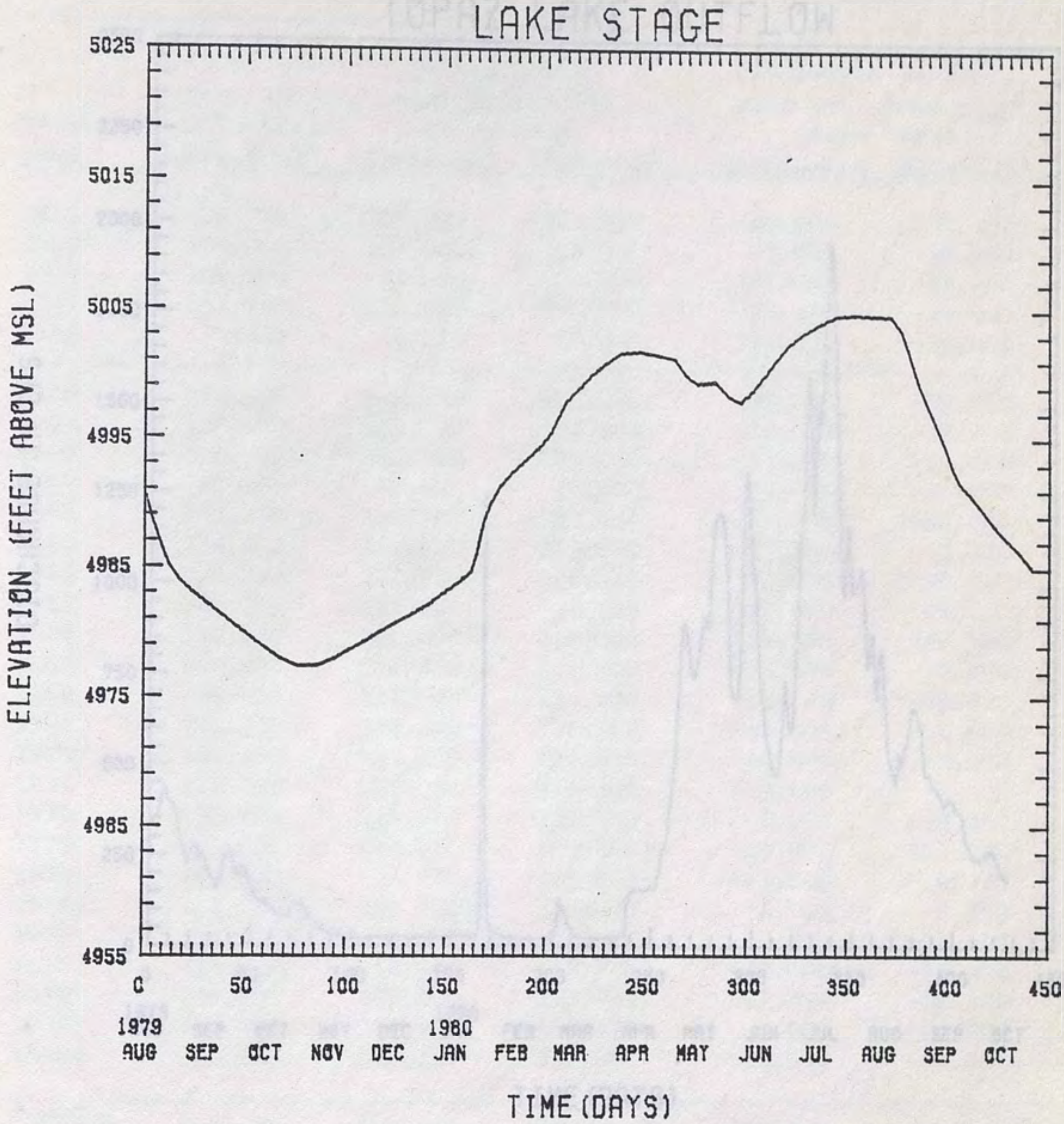


Figure 9. Topaz Lake Stage.

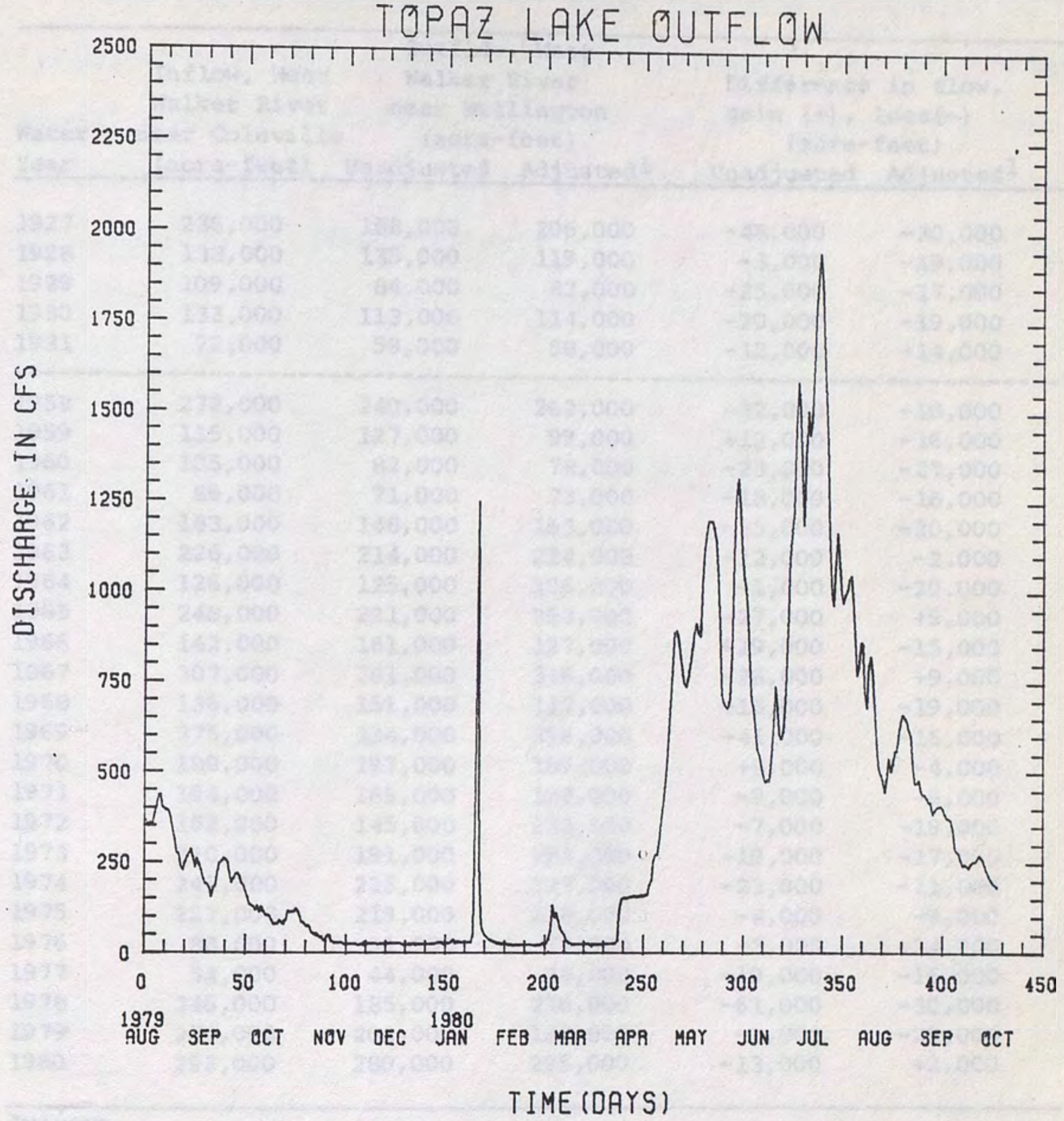


Figure 10. Topaz Lake Outflow.

Table 3.--West Walker River: Annual inflow, outflow and gain or loss, 1927-1980[2]

Water Year	Inflow, West Walker River near Coleville (acre-feet)	Outflow, West Walker River near Wellington (acre-feet) Unadjusted	Adjusted[1]	Difference in flow, gain (+), loss(-) (acre-feet) Unadjusted	Adjusted[1]
1927	236,000	188,000	206,000	-48,000	-30,000
1928	138,000	135,000	119,000	-3,000	-19,000
1929	109,000	84,000	82,000	-25,000	-27,000
1930	133,000	113,000	114,000	-20,000	-19,000
1931	72,000	59,000	58,000	-12,000	-14,000
1958	272,000	240,000	262,000	-32,000	-10,000
1959	115,000	127,000	99,000	+12,000	-16,000
1960	105,000	82,000	78,000	-23,000	-27,000
1961	89,000	71,000	73,000	-18,000	-16,000
1962	183,000	148,000	163,000	-35,000	-20,000
1963	226,000	214,000	224,000	-12,000	-2,000
1964	126,000	125,000	106,000	-1,000	-20,000
1965	248,000	221,000	253,000	-27,000	+5,000
1966	142,000	161,000	127,000	+19,000	-15,000
1967	307,000	281,000	316,000	-26,000	+9,000
1968	136,000	151,000	117,000	+15,000	-19,000
1969	375,000	334,000	358,000	-41,000	-16,000
1970	189,000	197,000	185,000	+8,000	-4,000
1971	194,000	185,000	188,000	-9,000	-6,000
1972	152,000	145,000	133,000	-7,000	-19,000
1973	210,000	191,000	193,000	-19,000	-17,000
1974	248,000	225,000	237,000	-23,000	-11,000
1975	227,000	219,000	218,000	-8,000	-9,000
1976	88,000	91,000	74,000	+3,000	-14,000
1977	54,000	44,000	38,000	-10,000	-16,000
1978	246,000	185,000	216,000	-61,000	-30,000
1979	204,000	202,000	182,000	-2,000	-22,000
1980	293,000	280,000	295,000	-13,000	+2,000
Average (Rounded)	183,000	168,000	168,000	-15,000	-14,000

1. Adjusted for storage changes in Topaz Lake.
2. Modified from Glancy, 1971.

of 1927-1931 and 1958-1980. Also on this table, the unadjusted and adjusted flows pertaining to changes in reservoir stage and volume in Topaz Lake are shown.

For the 28 year period of data, the average annual inflow past Coleville was 179,000 acre-feet, with an average annual adjusted outflow at Hoye Canyon of 164,000 acre-feet. This difference results in an average annual loss of 15,000 acre-feet in Antelope Valley.

TOPAZ LAKE

Topaz Lake is located in a topographically enclosed basin into which the unregulated flow of the West Walker River has been diverted. As irrigation demand increases during the summer and early fall months, gates in the outlet tunnel are opened to permit needed water to flow downstream to Smith and Mason Valley. These two factors, unregulated inflow and seasonally imposed control on outflow, produce fluctuating lake levels which are discernable from day to day (Figure 9), and a tidal effect on the adjacent groundwater system.

During the period of data collection, the lake stage fluctuated approximately 28 feet. The minimum stage was on October 20, 1979 when a stage of 4977.0 feet above sea level was reported. The maximum occurred on July 4, 1980 with a stage of 5005.5 feet above sea level. This change in elevation represents an increase in storage of 53,000 acre-feet during this time period (Figure 11, from Rush and Hill, 1972).

The surface area of the lake changes in response to these fluctuations because of a rather gradual sloping bottom. At the maximum stage of 5005.5 feet, the total surface area of the lake is 2400 acres and at the minimum stage of 4972.3, the area decreases to about 1500 acres (Figure 12, from Rush and Hill, 1972).

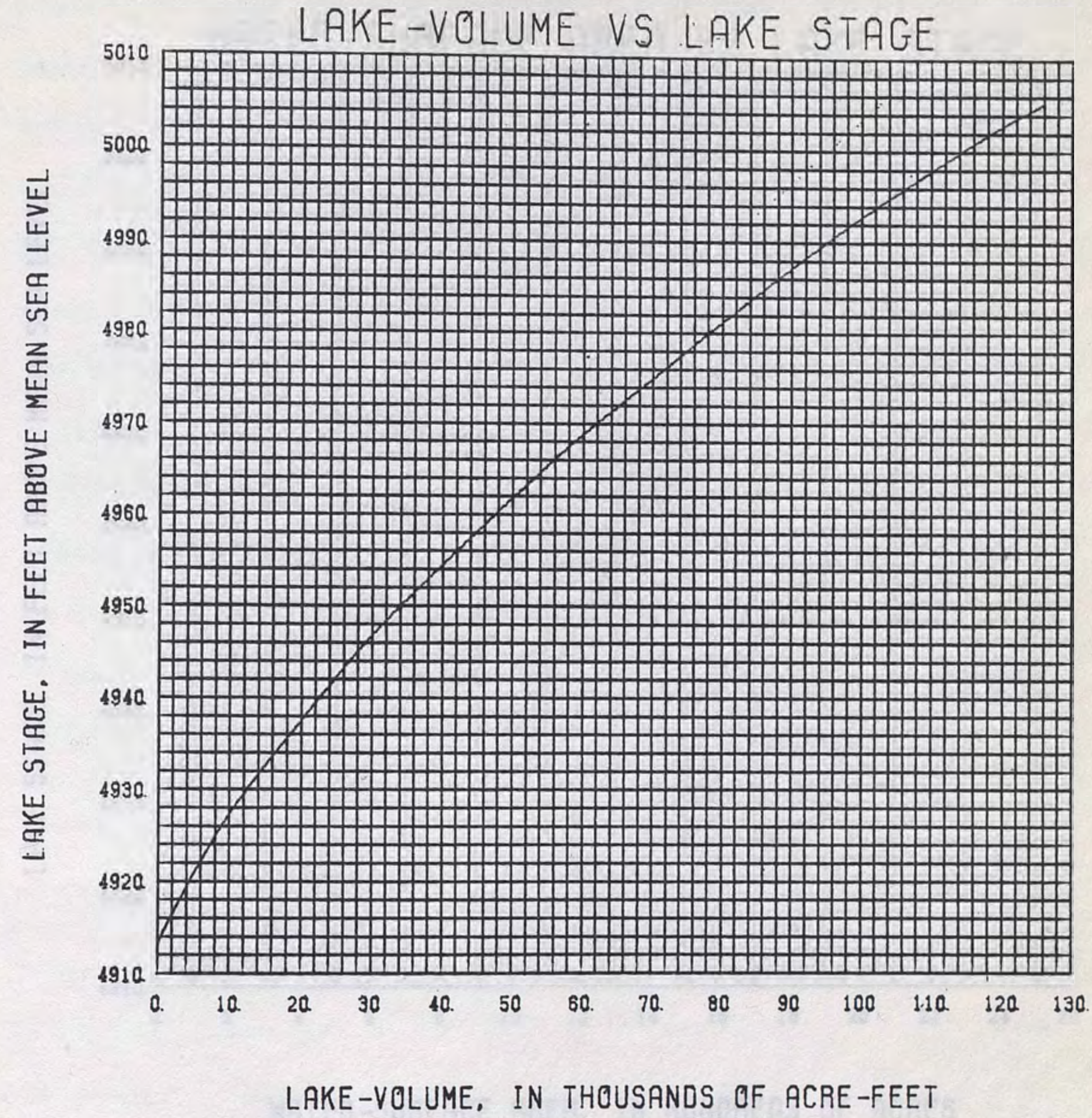


Figure 11. Topaz Lake: Volume versus Stage.

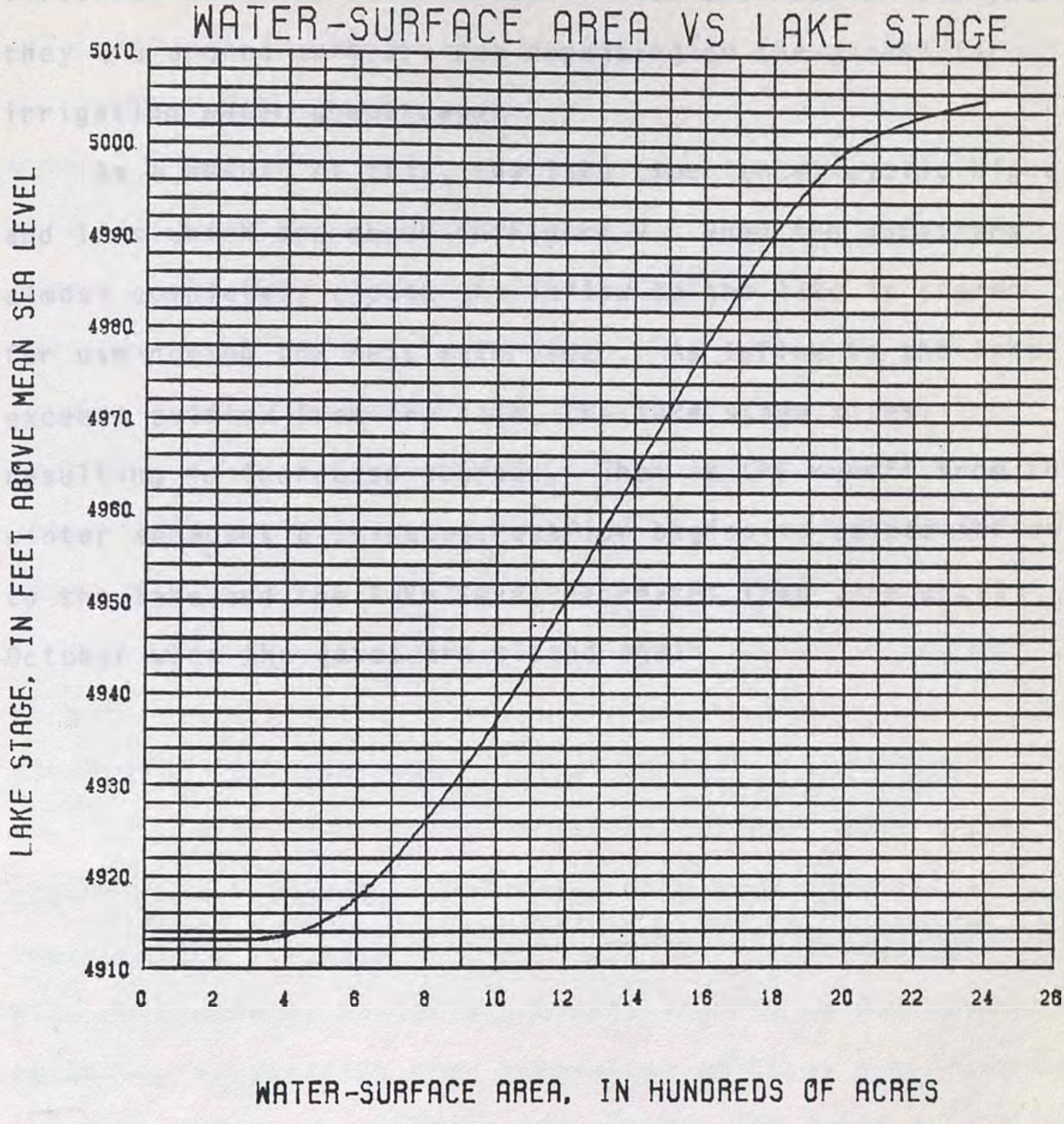


Figure 12. Topaz Lake: Area versus Stage.

The control gates on the outlet tunnel are governed by the W.R.I.D. and are seasonally opened and closed. From October to March, they are opened only enough to permit livestock water to pass through. Then the rest of the year, they are opened more or less depending on the demand for irrigation water downstream.

As a result of this, the lake experiences cyclic highs and lows which are shown in Figure 9. When the gates are almost completely closed the inflow to the lake is stored for use during the next water year. As inflow to the lake exceeds outflow from the lake, the lake stage rises, resulting in increased storage. Then as the runoff from the winter snowpack dissipates, outflow begins to exceed inflow to the lake and the lake level decreases from June until October when the gates are closed again.

EVAPOTRANSPIRATION

Most of the vegetation in Antelope Valley is found on the floodplains along the perennial streams. This vegetation is comprised mainly of phreatophytes and crops. The phreatophytes are deep-rooting plants which tap the groundwater to depths of less than about 50 feet, withdrawing water directly from the groundwater system. Some of the phreatophytes, like grasses, are of economic value, whereas others, such as greasewood and rabbitbrush, are of little economic importance. Work done in other areas by White (1932), Young and Blaney (1942), Houston (1950), and Robinson (1965), were utilized to obtain estimates of the consumptive-use rates in Antelope Valley shown in Table 4 (Glancy, 1971).

In Antelope Valley there is a vast number of perennial streams and irrigation ditches traversing the valley. This along with the free-water surface of Topaz Lake, makes this area have an unusually high total quantity of water evaporated from free-water surfaces. Upon examining past records from Lahontan Reservoir and Fallon, in conjunction with work done by Kohler and others (1959), an estimated value for evaporation from free-water surfaces was found to be about 4 feet per year in the valley. In Table 5, evaporation and evapotranspiration estimates are shown for the valley; these total 15,000 acre-feet per year.

Table 4.--Estimated Acreage and Consumptive Use by Crops in Antelope Valley[2]

Location of area	Crop	Estimated area (acres)	Estimated Consumptive Use: Rate (feet per year)	Estimated Consumptive Use: Acre-feet per year
		NEVADA		
North of West Walker River	Alfalfa[1]	400	2	800
Do.	Pasture[1]	160	2	320
Along West Walker River flood plain near State line	Pasture	a 280	2	560
		CALIFORNIA		
Antelope Valley	Alfalfa, pasture and grain	b 11,000	(c)	d 31,890
Do.	Native pasture (subirrigated by groundwater)	b 590	1.5	890
Total (rounded)		12,400		35,000

1. Irrigated mainly by pumping groundwater.
 a. C. N. Saulisberry, U.S. Dept. Agriculture, oral commun., 1969.
 b. California Dept. Water Resources, 1964, table 7,p. 66.
 c. Use rates from Calif. Dept. Water Resources (1964, table 10,p.81), as follows: alfalfa and pasture, 2.8 feet; grain(irrigated), 1.3 feet.
 d. California Dept. Water Resources , 1964, table 11, p. 82. Also includes 1,000 acre-feet of pumpage from wells drilled since compilation of the California report.

2. From Glancy, 1971.

Table 5.--Estimated Natural Evapotranspiration in Antelope Valley[1]

Area	General plant assemblage[2]	Estimated area (acres)	Approximate depth to water table (feet)	Estimated water use: Rate (feet per year)	Estimated water use: Acre-feet per year
		NEVADA			
West Walker River flood plain	Mostly native grass, willows, and other unidentified brush with some tules and associated marsh vegetation	1,500	0-20	1.5	2,200
North of West Walker River flood plain	Mostly greasewood and rabbitbrush with some scattered saltgrass	1,300	10-40	0.3	390
Do.	Mostly greasewood and some rabbit brush	440	20-50	0.1	40
Topaz Lake (part)	Free-water surface	1,080	0	a 4	4,300
West Walker River and irrigation drainage near State line	Do.	60	0	a 4	240
		CALIFORNIA			
Valley lowlands	Mostly native grass and willows; some cottonwood trees, greasewood, rabbitbrush, unidentified brush, and marsh vegetation	2,400	0-20	1.5	3,600
Topaz Lake (part)	Free-water surface	920	0	a 4	3,700
West Walker River and irrigation canals	Do.	100	0	a 4	400
Total (rounded)		7,800			15,000

1. Modified from Glancy, 1971.
2. Excludes native pasture, which is considered a crop.

Depth to water is generally greater than 30 feet throughout the alluvial fan area. The only area where evapotranspiration would occur would be around the littoral zone where the depth to water is less than 30 feet and phreatophytes are present. The areal extent of this zone is about 5 acres and would result in a loss of approximately one acre-foot per year of water to evapotranspiration. With these depths to water and such an insignificant zone where it could show influence, evapotranspiration was not accounted for in the study area.

GROUNDWATER

OCCURRENCE AND MOVEMENT

Most of the recoverable groundwater in the Topaz Lake region is found in the unconsolidated sediments of the alluvial fan. Where the alluvium consists of silts and clays, the permeability is low and only small well yields can be expected, with even lower well yields in the bedrock material. In contrast, the well sorted sand and gravel strata of buried stream channels in the fan, have moderate to high permeabilities and will yield water readily to wells intercepting them.

Water in the alluvial aquifer is unconfined and has its free-water surface at shallow depths, generally less than 200 feet below land surface. The depths to water range from zero at known springs to a depth of 250 feet in well #5 (Table 6). This range of depths is not as drastic as it seems for such a small area because the land surface here varies from about 5000 feet by the lake to around 5300 feet up along Highway 395.

The movement of groundwater follows a pattern similar to that of surface water; that is, groundwater flows from the topographically high regions towards the lower elevations of the valleys. In the alluvial aquifer of the study area, groundwater flow patterns were found to be dependent on lake stage, thus different from the normal patterns. Groundwater flow is generally considered to be from recharge

Table 6.--Well and Spring Data

Map no.	Location[1]	Name	Land[2] Surface Elev(ft)	Well Depth (ft)	Water Levels[4] Date	Depth (ft)
1	29 BACC1	Mann	5236[3]	Spg.	10-28-79	---
2	29 BCCB1	Davenpeck	5324[3]	Spg.	10-28-79	---
3	29 BBCD1	Fair I	5300	196	10-28-79	83
4	29 BBCD2	Fair II	5303	---	10-28-79	96
5	29 BDCB1	Clinton	5155	350	10-28-79	220
6	29 BDCA1	Breckenridge	5132	175	10-28-79	22
7	29 BDCD1	Busch	5113	185	10-28-79	144
8	29 BDDC1	Casale	5110	230	10-28-79	129
9	29 BDDD1	Brown I	5096	230	10-28-79	118
10	29 BDDD2	Brown II	5091	140	10-28-79	114
11	29 BDDA1	Freeland	5113	158	10-28-79	109
12	29 CACA1	Blankenship	5118	170	10-28-79	139
13	29 CACC1	Barnicle	5153	245	10-28-79	172
14	29 CBDB1	Librandi	5226	340	10-28-79	252
15	29 CCAA1	Topaz Lodge	5216	---	10-28-79	224
16	32 BBAD1	Pinenut	5127	206	10-28-79	152
17	32 BAAB2	McDaniels	5075	150	10-28-79	101
18	32 BAAC	---	5049	---	10-28-79	76
19	32 BADC2	---	5021	62	10-28-79	49
20	32 BADC1	Miller	5024	80	10-28-79	52
21	32 BADA1	Savage	5026	80	10-28-79	52
22	29 DCBA1	---	5020	---	10-28-79	46
23	29 DCBB3	Sherred	5039	---	10-28-79	64
24	29 DBCC	---	5055	---	10-28-79	79
25	29 CADD1	Rubino	5069	---	10-28-79	92
26	29 CADA1	Jones	5067	183	10-28-79	91
27	---	---	---	---	11-18-79	31
28	29 CDBB1	Mettler	5177	260	10-28-79	196
29	---	Nevada Creek	---	---	---	---

1. All wells are in T. 10 N. and R. 22 E., in Antelope Valley (hydrographic area 106). See text for well numbering system.
2. Determined with a surveying level. Datum is mean sea level. Accuracy ±1 foot.
3. From U.S.G.S. open file report #76-90.
4. Depth in feet below land-surface datum.

areas to discharge areas, moving downgradient in the direction of decreasing head. The groundwater system should show a water table mound under the higher elevations with its surface sloping down to the lake. Under these conditions, the groundwater will flow down to and be discharged at the lake.

After analyzing the system and data collected, the groundwater is found to move in this manner, but only when the lake level is lower than the adjacent water table. During this time, the gradient is sloping down to the lake, causing groundwater movement to be in the same direction. Evidence of this is shown in the plots of time versus well water-levels and lake level on the same graph (Appendix I). When the lake level begins to drop, the water levels in the wells drop a corresponding amount as a result of the draining of the aquifer to the lake. This reaction supports the idea of the lake being in hydraulic continuity with the groundwater system.

The lake stage typically reverses itself from steadily decreasing to increasing in stage during the fall, beginning towards the end of October. With this increase in stage, a gradient reversal occurs between the lake and the adjacent groundwater system, which causes the groundwater levels to increase a similar amount. This can be seen on the plots of lake stage versus well stage in Appendix I. These plots show that when the lake reaches a peak or trough, the water levels continue to rise or fall in the wells. They will continue to do so for a time which will be referred to as the phase lag. This phase lag is the time that it takes for a change in lake level to affect the water level in a given well.

When the gradient reversal occurs, the hydraulic head produced by the increased lake level results in the lake water seeping into the groundwater system, moving from a high to a low hydraulic head. This movement into the system continues as long as the lake level is rising resulting in an increase in the water table surface with time.

This increased head travels through the aquifer as a pulse at an average rate of 60 feet per day. At this rate, the wells along the shoreline, such as well #21, will show an increase in elevation within several days. The ones situated further up by Route 395 will not show any change from this pulse for approximately four weeks.

Once the pulse has propogated through the aquifer, the water table continues to rise as long as the lake does. While the lake is steadily rising, the water levels will exhibit a fairly constant rate of rise. A change will occur in this rate when the inflow to the lake has an extreme jump in flow. The resulting increase in storage and lake level will cause yet another pulse to form. Since the gradient has already affected the entire aquifer, the pulse travels as a second wave inducing a greater rate of change in the water levels as it moves through the aquifer. The gradient reversal remains established as long as the lake level is greater than the adjacent water table. Figure 13 is a graphical display of this phenomenon occurring in one of the wells. In comparing the inflow (Figure 13a) and outflow (Figure 13b) to Figure 13c, the cause and effect relationship between inflow/outflow and lake/well stage can be seen.

As the lake stage levels off to a stable elevation where inflow to the reservoir equals outflow from the reservoir, the hydraulic gradient reverses again. First, the water levels near the lake cease rising, then after corresponding phase lags have been met, water levels farther upslope will cease rising. At this time, the water table again exhibits a mound under the higher elevation areas and the gradient resumes its previous slope with flow back down to the lake.

This seasonal oscillation of water levels in both the lake and well, occurs at least twice ayear. The first time is in October when the outlet gates are closed to start storage for the following irrigation season. Then near the end of March, the gates are reopened in response to demand for irrigation water downstream. Although these are the principal occurrences of the gradient reversal, it may happen on a small scale if the direction of rise or fall is

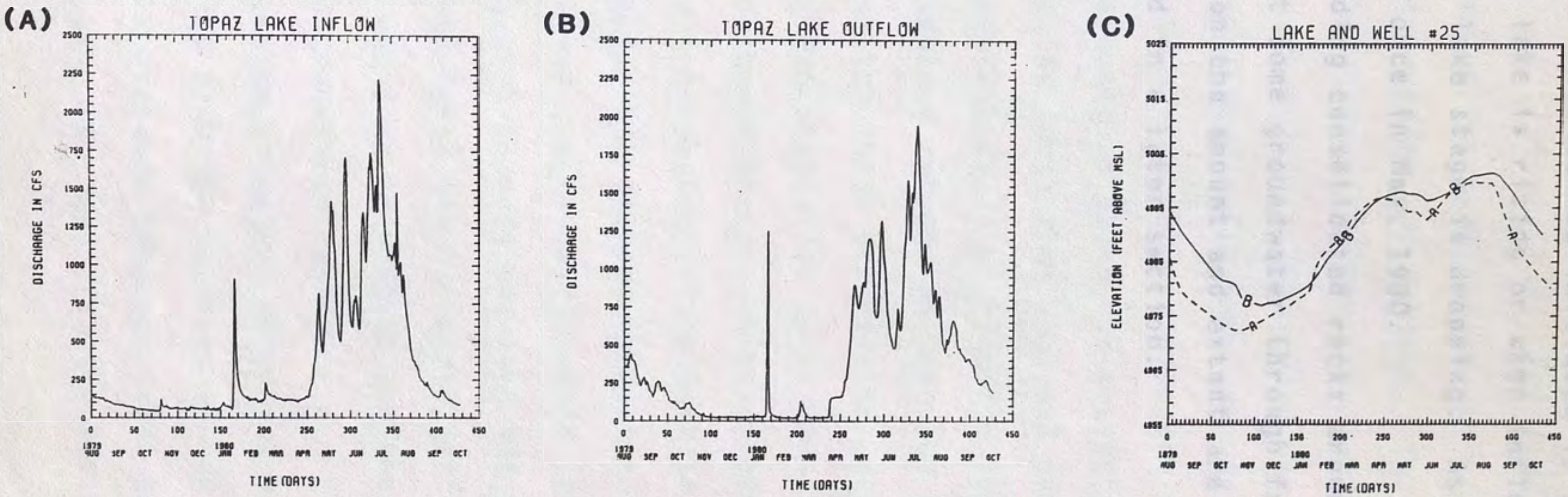


Figure 13. Relationship between A) Inflow, B) Outflow, and C) Lake/Well Stage.

suddenly altered. This can be initiated if outflow exceeds inflow while the lake is rising or when inflow surpasses outflow when the lake stage is dropping. As seen on Figure 9, this happened once in May, 1980.

The surrounding consolidated rocks around this alluvium probably transmit some groundwater through fractures as recharge. Data on the amount and extent are nonexistent and will be discussed in a later section.

SEEPAGE TO AND FROM TOPAZ LAKE

In the study by Winter (1976), the interaction between lakes and groundwater was examined in two dimensions and the idea of a stagnation point was defined. This represents the point of lowest head along the boundary between groundwater flow systems beneath a lake. This point was later found to be associated with a stagnation zone made up of a series of pseudostagnation points in three dimensions by Winter (1978). If the stagnation point was present, then the divide separating the local from regional or intermediate system would be continuous, isolating the local flow system, thus preventing leakage out of the lake and becoming the discharge point of the local groundwater system. In the other case where this stagnation point does not exist, the groundwater divide would be discontinuous and the lake would leak water to the groundwater system beneath the lake.

Another relationship deduced by Winter (1976) was that if the lake level was increased, it would tend to decrease the difference in head between the lake and the stagnation point. If the lake level was increased sufficiently, the stagnation point would be eliminated and the lake would then lose water to the groundwater system.

At Topaz Lake the situation which was of interest was not the interaction with the intermediate or regional flow systems, but rather between lake and local groundwater flow systems. By applying the idea of a stagnation point to the boundary of lake bottom sediments and the groundwater aquifer material, insight to the interactions of the lake and the adjacent groundwater systems can be obtained.

When the lake level decreases, the head difference between the groundwater and the lake system increases creating a stronger hydraulic gradient towards the lake. During this time the groundwater migrates towards the lower head of the lake, draining the aquifer of water. This continues as long as the lake level is lower than the adjacent water table.

The concept of a stagnation point on a boundary separating the lake system from the groundwater system seems to apply to this situation. As long as the lake level is decreasing, the boundary will have some minimum point on it which will be greater than the head represented by the lake surface, thus groundwater flow will be towards the lake.

Then, as the lake level rises, the head represented by the lake increases, decreasing the difference between the groundwater system and the lake. Once the lake head becomes greater than any piont on the boundary, the stagnation point is eliminated and the boundary becomes discontinuous. The break in the boundary results in a hydraulic gradient reversal so that the lake now loses water to the groundwater system. This water migrates into the aquifer towards lower head and is received as bank storage from the lake.

GROUNDWATER RECHARGE

Recharge of the Antelope Valley groundwater system is generally from precipitation within the hydrographic area. To estimate the amount of recharge which occurs, the method described by Eakin and others (1951, p. 79-81) was used by Glancy (1971). This method estimates the amount of recharge by calculating the amount of precipitation which falls in a certain altitude zone and multiplying it by a percentage, which yields a value for recharge in that zone.

On Table 7, the estimated precipitation and recharge potential for each zone are shown. Of the totals shown, the potential recharge from precipitation is about 10 percent which, according to Glancy (1971), is considerably lower than what has been computed for most of Nevada. Because of this large difference in values, it has been stated by Glancy (1971) that most of the runoff in this region becomes recharge for the consolidated rocks of the mountains.

In the alluvial fan area, groundwater recharge occurs from two sources. One is from the surrounding mountains, the other is from the reservoir. Data on recharge from fractures in the surrounding consolidated rocks are essentially non-existent but recharge is evident by the presence of springs on the northern and western contacts of the consolidated rocks and alluvium.

Recharge from the lake water occurs as bank storage during periods when the lake stage is greater than the

Table 7.--Estimated Average Annual Precipitation and Potential Groundwater Recharge[1]

Precipitation Zone (ft)	Area (acres)	Estimated Precipitation: Range	Estimated Precipitation: Ave. (ft)	Estimated Precipitation: Ave. (acre-ft)	Estimated Recharge: Percent of Precip.	Estimated Recharge: (acre-ft per year)
ANTELOPE VALLEY — Nevada Part						
Above 9,000	31	>24	2.0	62	25	16
8,000-9,000	2,010	20-24	1.8	3,600	20	720
7,000-8,000	9,300	15-20	1.5	14,000	15	2,100
6,000-7,000	17,200	12-15	1.1	19,000	7	1,300
5,000-6,000	36,900	8-12	.8	30,000	3	900
Below 5,000	4,810	<8	.5	2,400	--	--
California Part						
Above 9,000	4,180	>24	2.3	9,600	25	2,400
8,000-9,000	9,760	20-24	1.8	18,000	20	3,600
7,000-8,000	18,300	15-20	1.5	27,000	15	4,000
6,000-7,000	23,900	12-15	1.1	26,000	7	1,800
5,000-6,000	31,000	8-12	.8	25,000	3	750
Below 5,000	930	<8	.5	460	--	--
Total (Rounded)	158,000			175,000		18,000

1. From Glancy, 1971.

adjacent water table. During the study period, the lake was found to increase a total of about 28 feet, with the groundwater table rising anywhere from 28 feet near the shore to about 21 feet upgradient. Using an average of 25 feet, a specific yield of 0.10, and an area of about 0.3 square miles, the total amount of recharge to the groundwater system is about 500 acre-feet per year. This value represents the change in storage as a result of the increase in lake stage, and is a combination of recharge from the lake and the surrounding consolidated rocks. As the lake stage drops, this volume is then drained from the aquifer into the lake.

Indications of recharge from fractures in the consolidated rocks are found in the chemistry changes discussed in a later section and from the graphs of well level above lake level in Appendix II. An example of this is shown on Figure 14 where the water table increased 50 feet in about 30 days. This placed the head well above lake level for this time and it remained so for about 60 days. A possible explanation for this large rise in head is the existence of seasonal subsurface inflow to the alluvium from the consolidated rocks.

This movement of groundwater through fractures in the surrounding consolidated rocks of the basin-margin mountains into the alluvial aquifer of the valley basin, is referred to as hidden recharge (Feth, 1963). This water is derived from rain and snowmelt in the Sierra Nevada and moves as

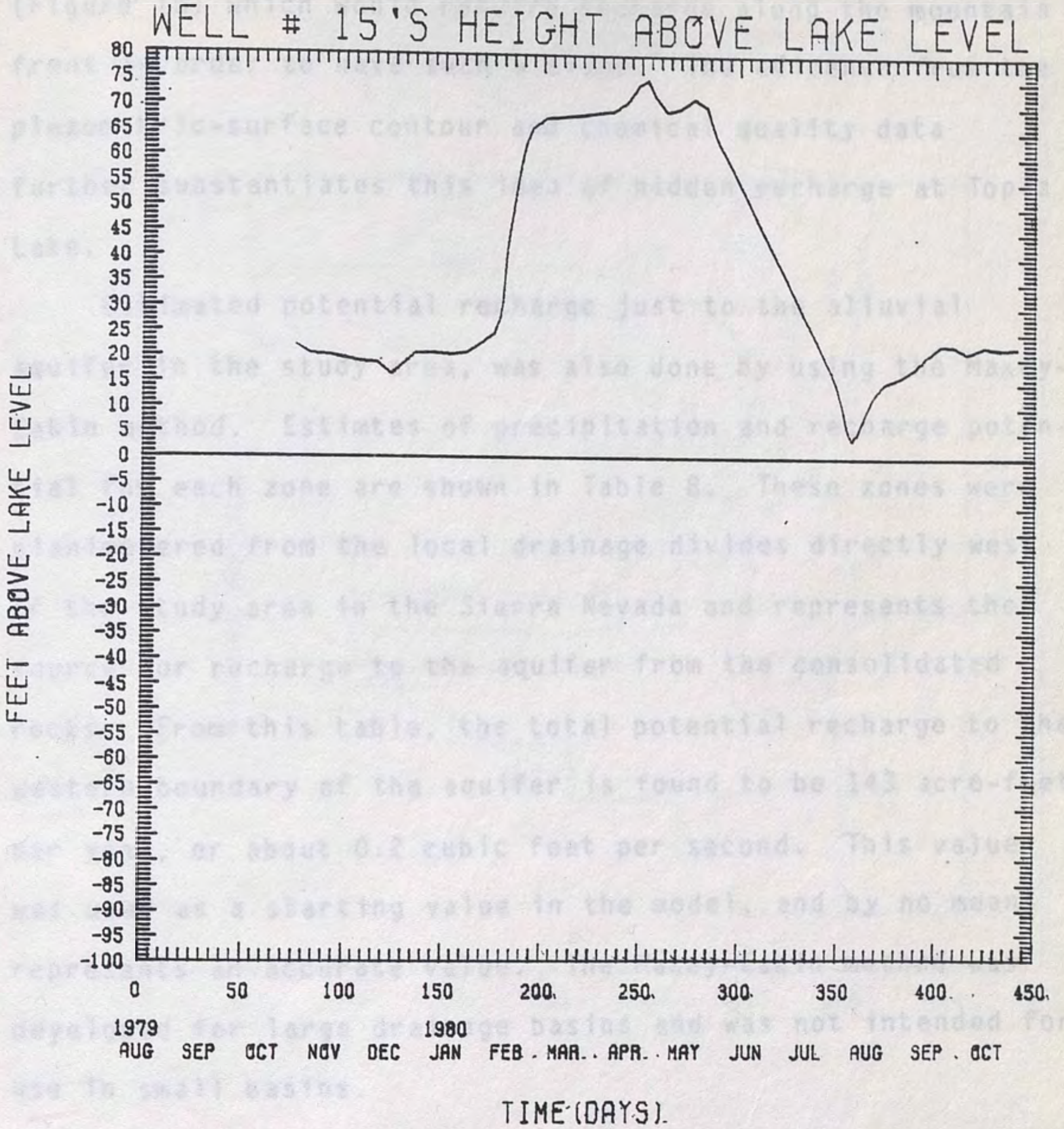


Figure 14. Well #15's Height Above Lake Level.

moves as subsurface percolation into the alluvium along the bedrock-alluvium contact. Further support of this concept is found on the configuration of the groundwater surface (Figure 15) which would require recharge along the mountain front in order to have such a slope. The evidence from the piezometric-surface contour and chemical quality data further substantiates this idea of hidden recharge at Topaz Lake.

Estimated potential recharge just to the alluvial aquifer in the study area, was also done by using the Maxey-Eakin method. Estimtes of precipitation and recharge potential for each zone are shown in Table 8. These zones were planimetered from the local drainage divides directly west of the study area in the Sierra Nevada and represents the source for recharge to the aquifer from the consolidated rocks. From this table, the total potential recharge to the western boundary of the aquifer is found to be 143 acre-feet per year, or about 0.2 cubic feet per second. This value was used as a starting value in the model, and by no means represents an accurate value. The Maxey-Eakin method was developed for large drainage basins and was not intended for use in small basins.

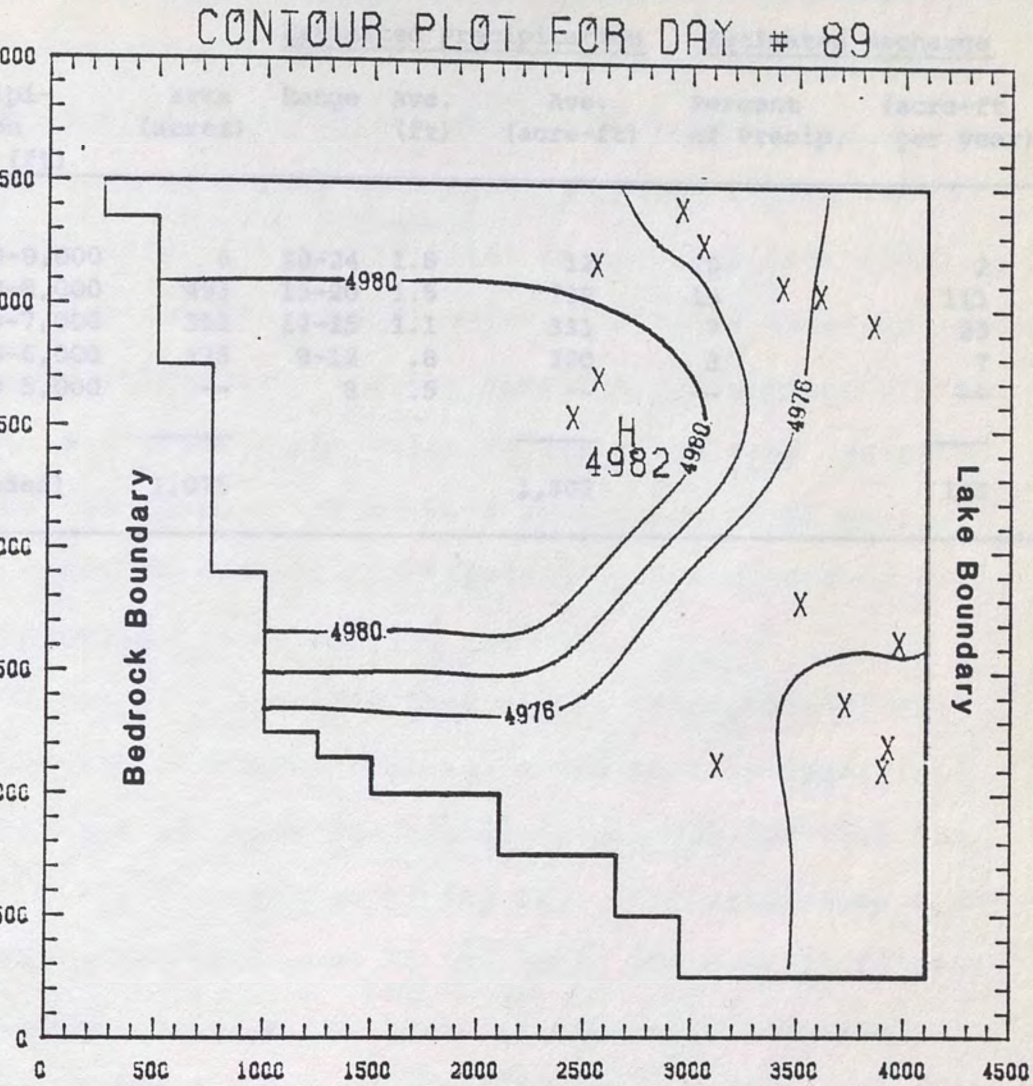


Figure 15. Contour Map of Piezometric Surface for Day #89.

Table 8.--Estimated Average Annual Precipitation and Potential Groundwater Recharge into the Modeled Zone

		Estimated Precipitation			Estimated Recharge	
Precipitation Zone (ft)	Area (acres)	Range	Ave. (ft)	Ave. (acre-ft)	Percent of Precip.	(acre-ft per year)
8,000-9,000	6	20-24	1.8	12	20	2
7,000-8,000	493	15-20	1.5	739	15	111
6,000-7,000	301	12-15	1.1	331	7	23
5,000-6,000	275	8-12	.8	220	3	7
Below 5,000	--	8	.5	--	--	--
Total (Rounded)	1,075			1,302		143

WATER QUALITY

GENERAL GROUNDWATER CHARACTER

As water moves through the subsurface it dissolves rock and soil constituents, thus increasing its dissolved solids concentration with increased contact or flow time. In view of this fact, waters which have been in the system for some time generally will exhibit higher values for electrical conductivity and total dissolved solids (TDS).

In chemical analyses of water, the value obtained for TDS can be calculated from the specific conductance of the sample. Analyses of the water in the study area indicated that the TDS concentrations were an average of 82 percent of the specific conductance (Table 9). The method of calculating these values for TDS is:

TDS (mg/l) = Specific Conductance (micromhos)X(.82)

Analyses of waters from wells and springs upgradient from the lake all show the higher values for TDS with the highest being 655 mg/l at spring #2. This value then tends to steadily decrease down to the lake, where wells closest to the shoreline have the lower values, with the lake having the lowest value of about 103 mg/l. This indicates the possibility of a mixing of two chemically different waters.

Classification of water can be done on the basis of the prevailing cations and anions present expressed as a relative percentage of the total number of cations and

Table 9.--Chemical Analyses of Ground and Surface Waters

Sample map#	Date	SiO_2 mg/l	Na^+ mg/l	K^+ mg/l	Ca^{2+} mg/l	Mg^{2+} mg/l	HCO^- mg/l	Cl^- mg/l	SO_4^{2-} mg/l	NO_3^{-}* mg/l	EC µmhos	pH (lab)	TDS mg/l	Temp. °C
1	07-05-79	30.0	23.1	..45	58.5	8.7	244.	3.5	22.0	.11	405	6.88	360	18.0
2	08-01-79	29.2	22.8	1.20	114.0	25.3	301.	2.8	1.8	6.77	770	7.55	655	23.5
3	08-01-79	46.0	19.4	.99	36.0	10.4	128.	2.6	57.5	1.64	332	7.10	256	17.0
4	11-27-79	53.5	21.9	1.12	32.8	7.4	131.	3.5	52.2	.09	301	7.02	250	13.5
5	08-01-79	36.0	16.8	.73	44.3	8.64	172.	4.5	37.5	.09	348	7.93	247	20.0
6	08-01-79	33.1	20.2	.66	46.6	9.81	204.	4.0	30.0	.04	378	7.78	315	18.5
7	10-16-79	32.5	22.4	..75	51.8	8.45	198.	6.8	24.0	23.70	402	7.67	335	17.0
9	08-01-79	45.0	15.4	.23	38.4	5.75	139.	7.8	12.7	22.00	298	7.40	241	18.5
11	08-15-79	39.0	24.3	.20	50.5	7.30	202.	3.5	30.2	1.24	380	7.20	319	17.5
12	08-01-79	35.1	17.8	2.40	43.0	9.71	154.	2.5	46.0	15.10	355	7.45	290	17.0
13	08-15-79	34.0	13.8	2.50	43.5	8.20	153.	1.0	42.0	3.20	321	7.02	267	15.0
15	10-01-79	49.0	9.8	3.80	31.2	9.45	146.	4.2	12.6	7.30	271	7.20	224	15.0
16	08-01-79	41.5	7.9	3.20	21.8	8.00	118.	1.2	9.2	.15	198	7.20	169	11.5
17	10-01-79	52.0	13.0	5.40	51.5	17.00	236.	12.0	13.3	12.30	432	7.27	360	16.0
	07-22-80	51.0	12.3	5.30	44.7	14.60	212.	10.8	11.7	10.20	384	7.11	321	----
20	10-01-79	45.0	8.0	3.40	21.2	6.80	109.	2.8	9.6	.66	190	7.32	161	15.0
21	08-15-79	55.0	10.8	4.50	28.5	8.20	143.	.5	13.0	2.30	247	7.11	211	17.0
	07-22-80	56.0	9.8	4.40	26.0	7.87	139.	1.4	10.9	1.88	236	7.23	215	----
23	08-01-79	50.7	10.6	3.66	28.6	9.87	139.	1.0	22.5	1.82	262	7.46	217	18.0
	07-22-80	51.0	11.2	4.10	27.3	9.56	140.	1.3	20.2	1.89	264	7.09	215	----
25	08-01-79	36.0	15.3	2.51	48.2	11.10	157.	2.3	65.6	5.50	382	7.70	307	18.0
26	10-01-79	40.0	22.5	1.50	62.5	16.40	200.	3.8	100.0	14.00	508	7.28	420	19.0
A	09-02-80	18.0	13.4	3.50	16.6	5.20	108.	3.4	7.0	-----	197	7.90	158	----
29	08-01-79	44.3	8.1	2.50	19.0	5.95	104.	.7	7.2	.15	164	7.55	147	18.5
	12-05-79	39.0	7.5	3.00	15.4	5.32	85.8	.9	7.8	.40	152	7.82	126	----
Lake	08-07-79	6.9	9.1	1.70	12.5	3.10	69.0	3.2	4.8	.34	135	7.31	104	----

* NO_3^- >10 mg/l may exhibit septic tank influence.

anions by using a trilinear diagram. With the use of this diagram, the chemical composition and homogeneity of the samples can be evaluated by plotting on it the relative percentages of the major cations and anions present. Waters representing a mixture of the two will manifest themselves as a series of points aligned between the end points. One other useful facet of this diagram is that if the plotting points line up in a direction pointing to a vertex, it represents a change in the relative percent of ions depicted by that vertex.

In Figure 16, the subfield to the left shows that the predominant cation in all the waters analyzed was calcium. The lower concentrations were found to be in the wells closer to the lake with the lowest, 53 percent, at well #23. This value then increases upgradient to a maximum of 70 percent at spring 2. The relative percentage of magnesium is less than 20 percent in all the samples with no discernable pattern.

Sodium plus potassium percentages also show a trend in the analyses with respect to location. Samples from wells near the shoreline have the highest relative percentage and upgradient points have decreasing values. The range of values varies from about 15 percent at spring 2 to a high of 30 percent at well #21, with the lake having 39 percent.

Upon examining Figure 16, a linear trend is exhibited on the right subfield representing the anions. The trend is

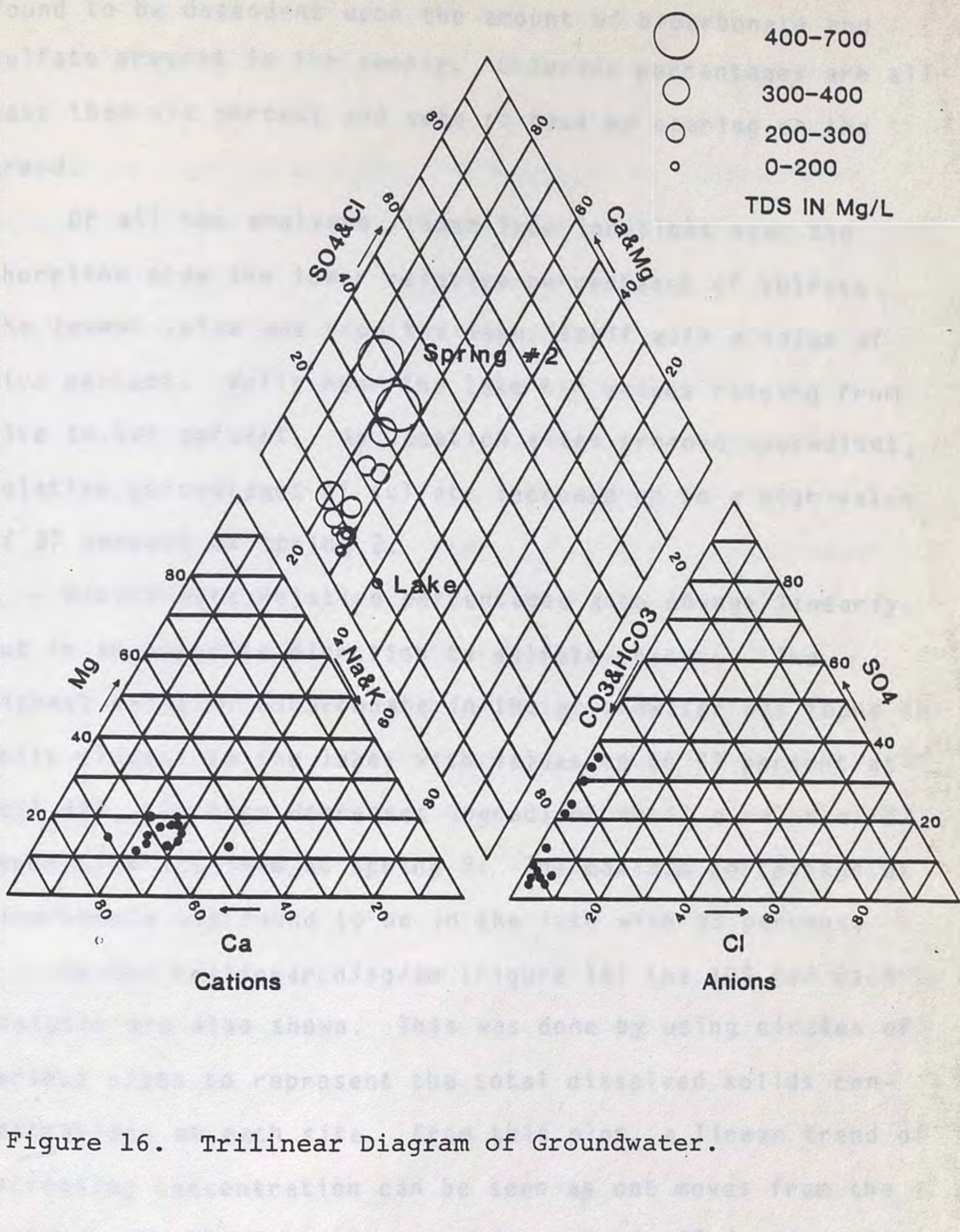


Figure 16. Trilinear Diagram of Groundwater.

found to be dependent upon the amount of bicarbonate and sulfate present in the sample. Chloride percentages are all less than six percent and seem to have no bearing on the trend.

Of all the analyses, those from locations near the shoreline show the lower relative percentages of sulfate. The lowest value was from the lake itself with a value of five percent. Wells near the lake had values ranging from five to ten percent. As location sites proceed upgradient, relative percentages of sulfate increase up to a high value of 37 percent at spring 2.

Bicarbonate relative percentages also change linearly, but in an opposite direction to sulfate changes. The highest value of bicarbonate in the groundwater was found in wells closest to the lake, with values up to 92 percent at well #21. It then decreases upgradient until a value of 62 percent is attained at spring 2. The maximum percentage of bicarbonate was found to be in the lake with 93 percent.

On the trilinear diagram (Figure 16) the TDS for each analysis are also shown. This was done by using circles of various sizes to represent the total dissolved solids concentrations at each site. From this plot, a linear trend of increasing concentration can be seen as one moves from the lake site towards analysis #2 (spring site) in an upgradient direction. This direction is opposite to what one would expect since older waters with higher TDS should be found downgradient.

The extent to which lake water migrates into the groundwater system can be deduced from these changes in water chemistry. The farthest upgradient water, spring 2, is the only water analyzed which showed no changes from mixing with a water with a chemistry similar to that of the lake water. All other analyses show a shift in their chemistries caused by the mixing of water similar to the spring sample and lake water. This is reflected in the shift of relative percent of certain ions and anions as sample sites progress from the lake to the spring area.

From Figure 16 the concept that there are two water sources which have mixed is further supported. These two waters represent either ones from upgradient sites, low bicarbonate percentage and high TDS (analysis #2), or from surface water, high bicarbonate and low TDS (lake analysis). The upgradient waters have the highest TDS concentrations and the surface waters have the lowest concentrations with varying concentrations found in waters in between these two sources. These sites located in between the two end points, lake and spring water, represent mixtures of upgradient and surface waters.

SURFACE WATER

Lake water quality throughout the study period varied little. Plotted on a trilinear diagram, Figure 17 monthly analyses all fell into the same general area with little variation in relative percentages of cations and anions.

Of the anions, bicarbonate was in the greatest quantity, 93 percent of total anions, with chloride and sulfate both around three percent. In the cations calcium and potassium plus sodium, the ranges were 44 to 49 percent and 39 to 44 percent respectively. With these percentages, the lake water can be characterized as a bicarbonate water, low in chloride and sulfate, and not having any predominant cation. Nevada Creek also exhibits similar chemical character but with slightly higher calcium percentage.

Fluctuations in lake inflow and outflow (Figures 8 and 10), and therefore volume (Figure 18), are reflected in the varying conductivity and TDS measured in the lake waters. As the lake is increasing in stage, from October when the gates are closed to March when the gates are opened again to release water, the reservoir acts as a collection tank with no outlet. The detention time in such a closed-basin reservoir creates an environment where the continued influx of solutes and surface evaporation tends to increase both TDS and electrical conductivity.

When water is released from March through July, the inflow to the lake is still greater than outflow, therefore

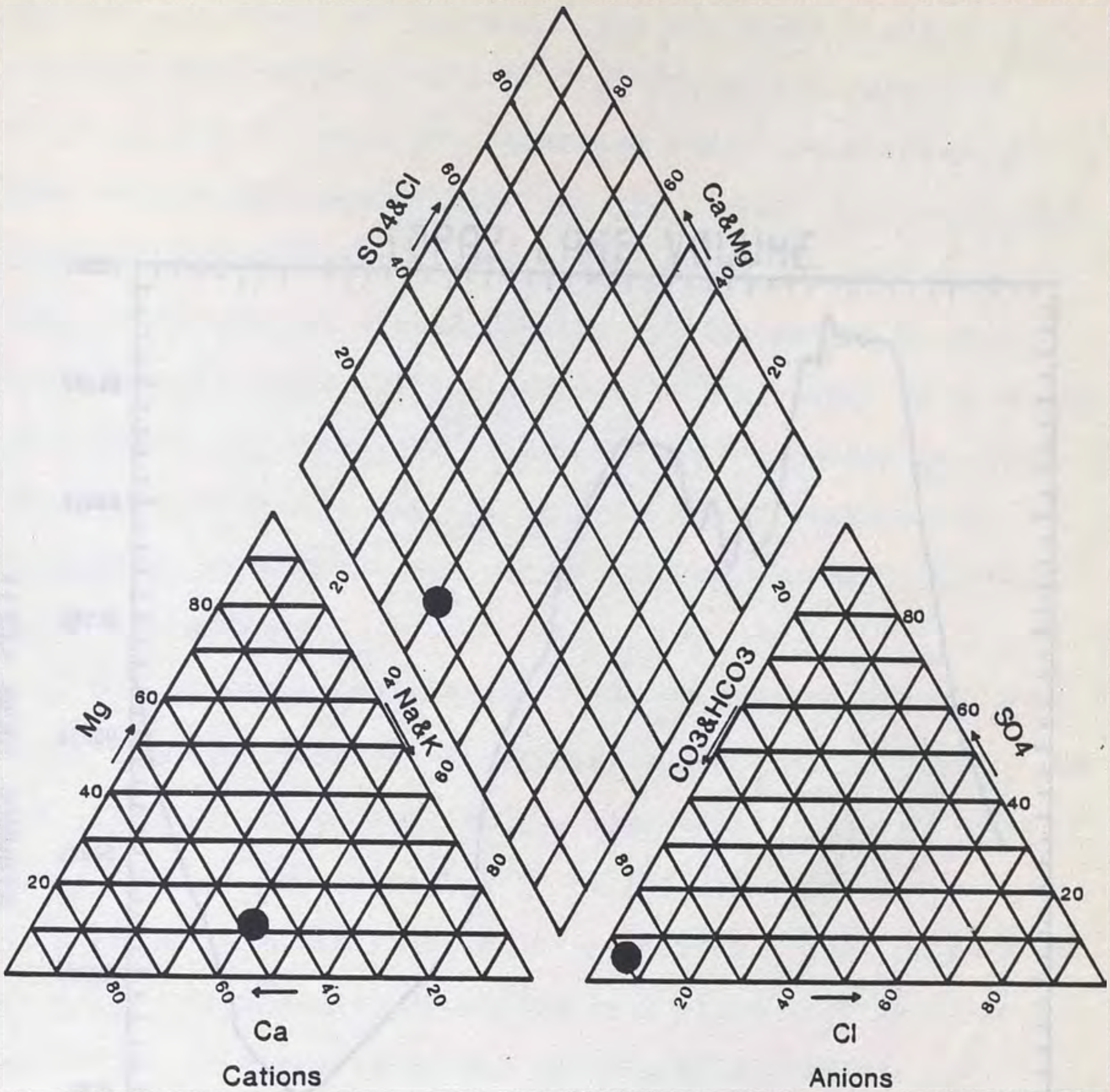


Figure 17. Trilinear Diagram of Surface Water.

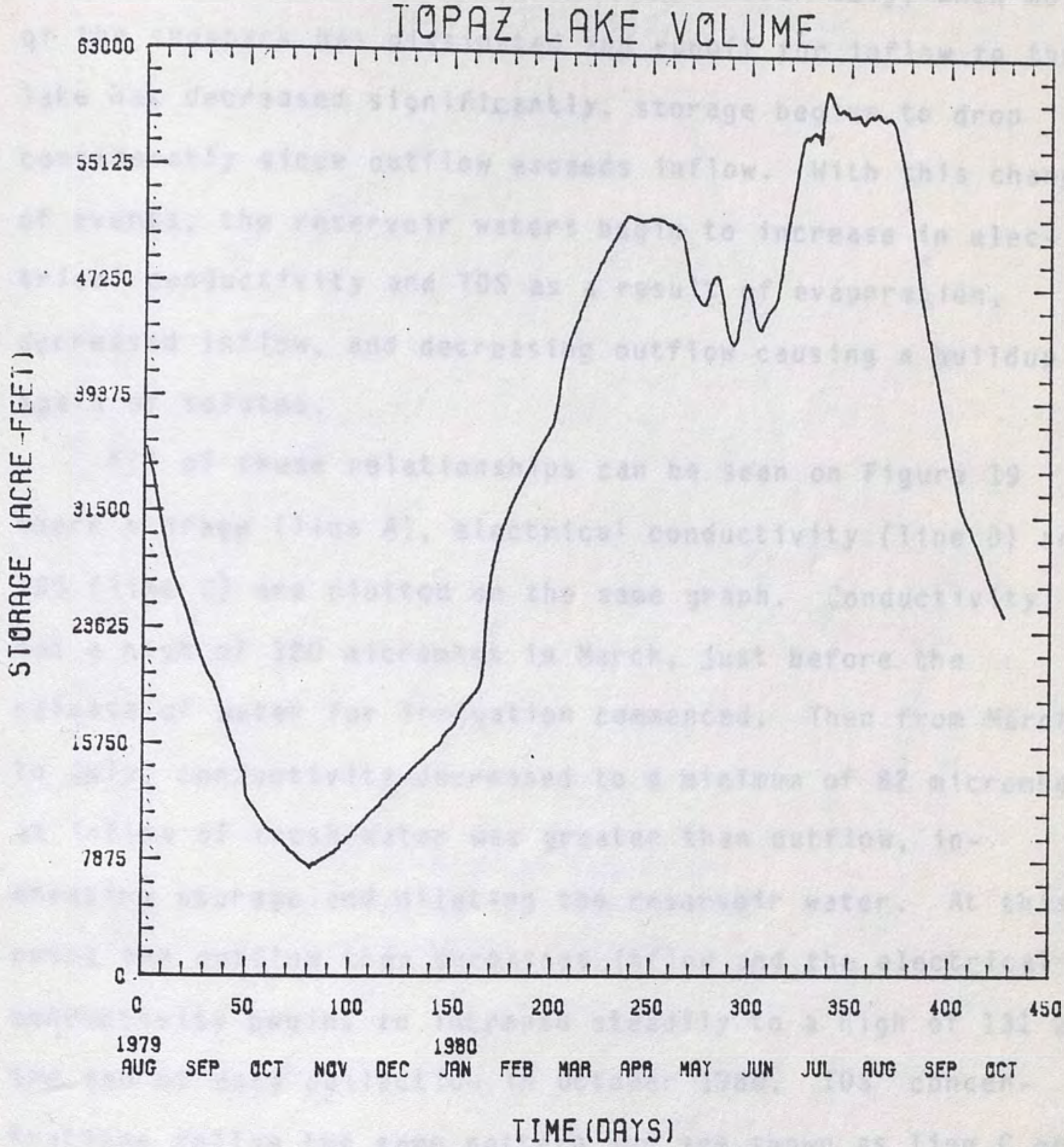


Figure 18. Topaz Lake Volume.

storage continues to increase. But now since there is some outflow, the incoming waters move through the reservoir flushing and diluting the reservoir water and decreasing both electrical conductivity and TDS. After July, when most of the snowpack has dissipated and runoff for inflow to the lake has decreased significantly, storage begins to drop considerably since outflow exceeds inflow. With this change of events, the reservoir waters begin to increase in electrical conductivity and TDS as a result of evaporation, decreased inflow, and decreasing outflow causing a buildup again of solutes.

All of these relationships can be seen on Figure 19 where storage (line A), electrical conductivity (line B) and TDS (line C) are plotted on the same graph. Conductivity had a high of 180 micromhos in March, just before the release of water for irrigation commenced. Then from March to July, conductivity decreased to a minimum of 82 micromhos as inflow of fresh water was greater than outflow, increasing storage and diluting the reservoir water. At this point the outflow then surpasses inflow and the electrical conductivity begins to increase steadily to a high of 131 at the end of data collection in October 1980. TDS concentrations follow the same pattern and are shown as line C on Figure 19.

Nevada Creek, which is an ephemeral stream transversing the alluvial fan on the southern portion of the study area

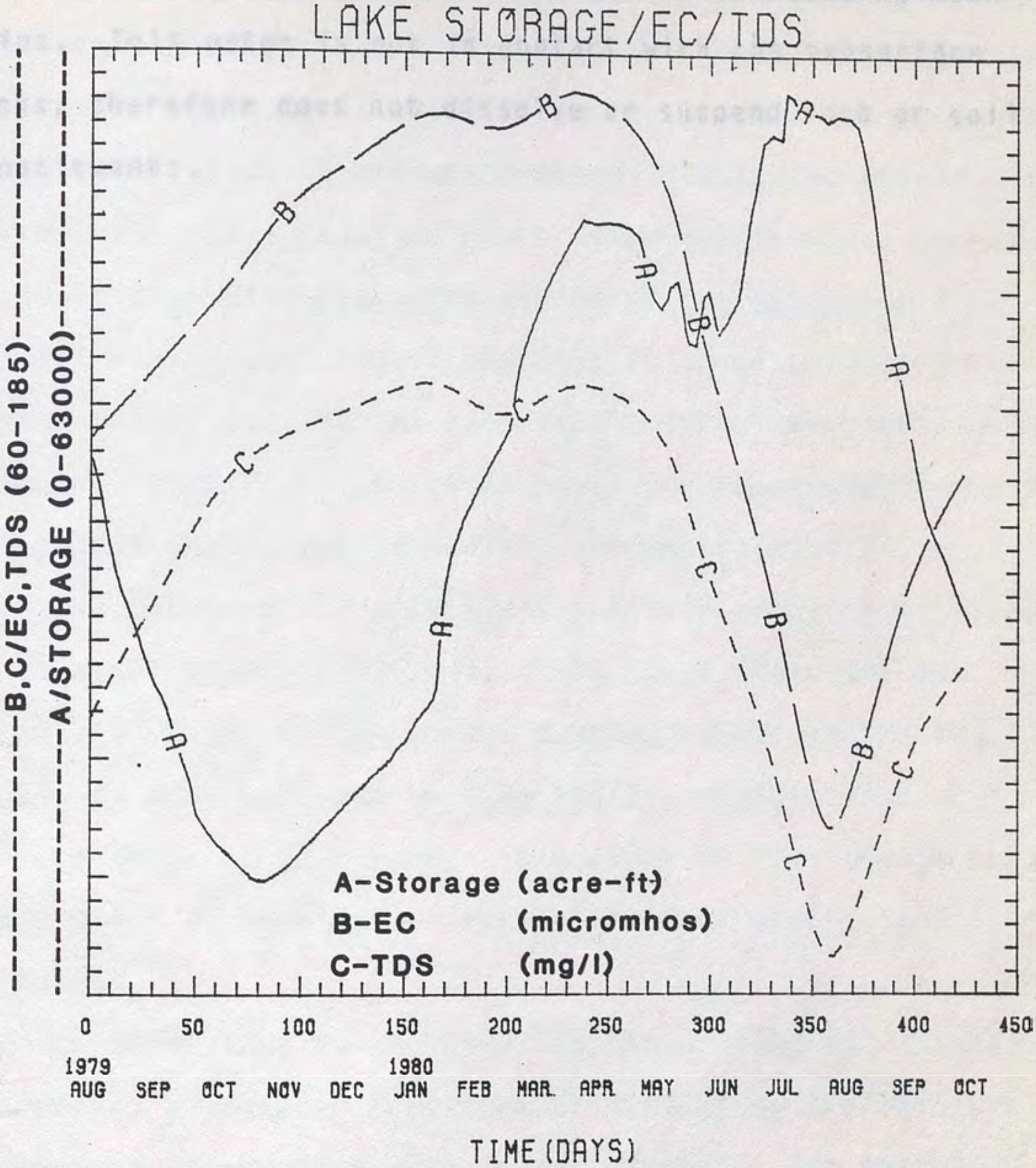


Figure 19. Superposition of Topaz Lake Storage, Conductivity, and Dissolved Solids.

had an average conductivity of 158 micromhos, and an average TDS of 137 mg/l. These low values for both parameters are due to the fact that water supplying the creek is derived mainly from runoff waters in the surrounding mountains. This water is not in contact with the subsurface rocks, therefore does not dissolve or suspend rock or soil constituents.

SANITARY CONDITIONS OF THE GROUNDWATER

When examining a water supply for domestic use, most considerations are aesthetic in nature and will not mark a water as unsuitable for human consumption, however, there are two parameters that if found in high concentrations, can render the water unsuitable for domestic use. The first is the concentration of nitrate present. Contamination from septic tanks or irrigation return flow is often the source of these high nitrates. The second is the existence of harmful microorganisms. These usually occur jointly with high nitrate, but one can be present without the other. The standard concentrations listed below are ones established by the United States Public Health Service (1962).

The standard for acceptable concentrations of nitrates in domestic water is 45 mg/l. This upper limit was not surpassed by any of the waters sampled. Well #3 had the lowest at 0.04 mg/l and well #6 had the highest with a concentration of 23.7 mg/l. This value is high enough that there might be some local degradation from septic tank influence.

In attempting to quantify the bacteriological quality of waters, a standard procedure of evaluating the coliform bacteria concentration present was adopted. The common species Escherichia coli is used as an indicator of possible fecal pollution. This species is not harmful itself, but since it is found in the intestinal tracts of warm-blooded animals, its presence along with related bacteria can be indicative of other species which may be pathogenic.

On April 4, 1980, a sample collection was done at 12 wells and two lake surface locations in order to evaluate coliform concentrations. In Table 2, the concentrations found are expressed as coliform colonies per 100 milliliters of sample. These results show that no coliform colonies were present in any of the wells. This is in compliance for meeting the United States Public Health Service standards on drinking water (U.S.P.H.S., 1962). Analyses of the surface waters showed the presence of one and seven colonies present in 100 milliliters of sample, well below the standard set for bathing waters, which is less than 200 colonies per 100 milliliters of sample (U.S.E.P.A., 1976). To be totally assured of no bacteria contamination, multiple samples should be collected over a period of several days at regular intervals.

SPECIFIC CONDUCTIVITY CHANGES

For a period of about one month while the lake level was dropping, a specific conductivity survey of water seeping into the lake was undertaken on a weekly basis at three sites along the shoreline. This was done during the month of September when the groundwater gradient was known to be sloping towards the lake with the water table declining an average of about 1.75 feet per week.

Since the seepage was infiltrated lake water having a lower specific conductivity, the mixing of it with the groundwater of the aquifer, resulted in water with a lower electrical conductivity. When drainage of the aquifer occurs, upgradient water with a higher conductivity flushes the system, thus increasing the conductivity of downgradient zones with respect to time.

The results of the study showed that with time the electrical conductivity was increased. Lake waters at this time had a low conductivity reading of about 115 micromhos which increased only to 131 micromhos at the end of the month. This increase represents a 14 percent rise in conductivity. Site A on Figure 2 had conductivity readings increasing from 197 to 337 micromhos during the same time span, signifying an increase of 70 percent. Site C had values increase from 127 to 147 micromhos, showing an increase of about 16 percent for the same period. Site B was in place only for the first measurement because the meter was damaged between the first and second week. The sample collected showed a conductivity of 128 micromhos which was the same as site C. These lower values from B and C seem to reflect some recharge from a recharge zone under Nevada Creek which showed a low value for conductivity.

During this span wells adjacent to the shoreline also showed increases in conductivities with respect to time. Well #23 showed an increase from 240 to 380 micromhos which is an increase of 58 percent.

SUMMARY

In discussing groundwater chemistry, there is a general trend which usually is found to hold true. With increased contact and flow time, TDS (and thus specific conductivities) should increase downgradient in the system. Both of these trends arise from the dissolution of rock and soil constituents as the water travels through the subsurface material.

Upon examining the analyses, it is seen that neither of these trends holds true at Topaz Lake; instead just the reverse is found. Closer to the bedrock-alluvium contact conductivity and TDS are the highest. With the high TDS, both sulfate and calcium are present in high concentrations. This represents groundwater being recharged by water similar in chemistry to that of the spring.

Proceeding downgradient specific conductivity and TDS both decrease until the lowest value is encountered in the lake water. The water chemistry of these samples progressively changes downgradient from one similar to spring #2 to one similar to the lake water chemistry. Wells located in between the two represent a transition zone between spring and lake chemistries caused by the mixing of the lake water as bank storage during periods of high lake stage.

Upon examining the results of the bacteriological investigations, neither high nitrate concentrations nor the presence of coliform bacilli detrimental to human health were found in any of the samples at levels high enough to be of concern. At this point in time the septic tanks do not seem to have contaminated the groundwater system on a whole. Areas showing elevated nitrate levels, which have been so over the last few years, should be the object of a localized study of the septic tank/well systems looking for a localized degradation problem.

MODEL SIMULATION

COMPUTER MODEL

The model used to describe groundwater flow in this thesis, was the finite difference model (FDM) for aquifer simulation in two dimensions (Trescott, Pinder, and Larson, 1976). It was designed to simulate an aquifer's response to various imposed stresses, whether it be artesian, water table, or a combination of the two. The model is extremely flexible in that it can simulate a heterogeneous and anisotropic system with irregular boundaries, and it can consider leakage, constant recharge, constant head, evapotranspiration, and well discharge within the aquifer.

The documentation and input required for the program is rather straightforward and easy to understand. The program used was originally written for an IBM 370/155 computer using FORTRAN IVG, but since the computer available was a CDC Cyber 171, a CDC version was obtained from the U.S.G.S. This version was written in FORTRAN IV language and required minor modifications to run on the Cyber 171. The most important modification to the program was to incorporate a term which would account for the fluctuating head boundary of the lake surface and is discussed in the Boundary Condition section.

The use of computer groundwater flow models as predictive tools for groundwater systems is having a great impact on groundwater resource development and planning.

The rationale behind the increased usage of these predictive models is that if they can reproduce historical hydrologic data, they can be used to approximate future groundwater conditions of the area under various utilizations.

During calibration procedures, geohydrologic data of the flow system was being incorporated into the model. Most of the time spent in development of the program was done in calibrating the model to reproduce the historic data collected. This calibration was done by adjusting input parameters, within reasonable real world limits, until a close approximation of historic data was reached.

MATHEMATICAL DESCRIPTION

The model used for the aquifer adjacent to Topaz Lake implements the iterative ADI procedure of solving the finite-difference approximations of the nonlinear, partial differential equations which describe nonsteady two-dimensional groundwater flow. The ADI procedure is a numerical technique used to solve sets of simultaneous equations derived from the finite-difference approximations described by Peaceman and Rachford (1955).

The general equation which describes two-dimensional groundwater flow in a confined aquifer can be written as

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial x}\left(T_{xy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial y}\left(T_{yx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right) = S\frac{\partial h}{\partial t} + W(x,y,t) \qquad (1)$$

where

T_{xx}, T_{xy}, T_{yx}, T_{yy}	are the transmissivity tensor components (L^2/t);
h	is the hydraulic head (L);
S	is the storage coefficient (dimensionless)
$W(x,y,t)$	is the volumetric flux (recharge or withdrawal) per unit area (L/t).

Further discussion and derivation of equation 1 can be found in Pinder and Bredehoeft (1968).

To simplify equation 1, one can assume that the Cartesian coordinate axes x and y are aligned with the principal axes of the hydraulic conductivity tensor. This simplification gives the equation

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right) = S\frac{\partial h}{\partial t} + W\,(x,y,t). \qquad (2)$$

In an unconfined aquifer, the transmissivity is a function of the saturated thickness of the aquifer, which in turn is a function of head. Equation 2 now becomes (Bredehoeft and Pinder, 1970)

$$\frac{\partial}{\partial x}\left(K_x\, b\,\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y b\,\frac{\partial h}{\partial y}\right) = S_y\frac{\partial h}{\partial t} + W(x,y,t) \qquad (3)$$

where

K_x, K_y are the principal components of the hydraulic conductivity tensor (L/t);

Sy is the specific yield of the aquifer (dimensionless);

b is the saturated thickness of the aquifer as a function of head (L).

To solve equation 2, a rectangular grid is superimposed over the study ara, each rectangle is assumed to possess uniform aquifer properties throughout (Figure 20). Hydraulic parameters, such as hydraulic conductivity, saturated thickness, water levels, and storage were either calculated, measured, or estimated for the center of each block. Each block uses a node at the center of the block as the point at which the continuous derivatives of equation 2 are approximated by finite-difference expressions. There

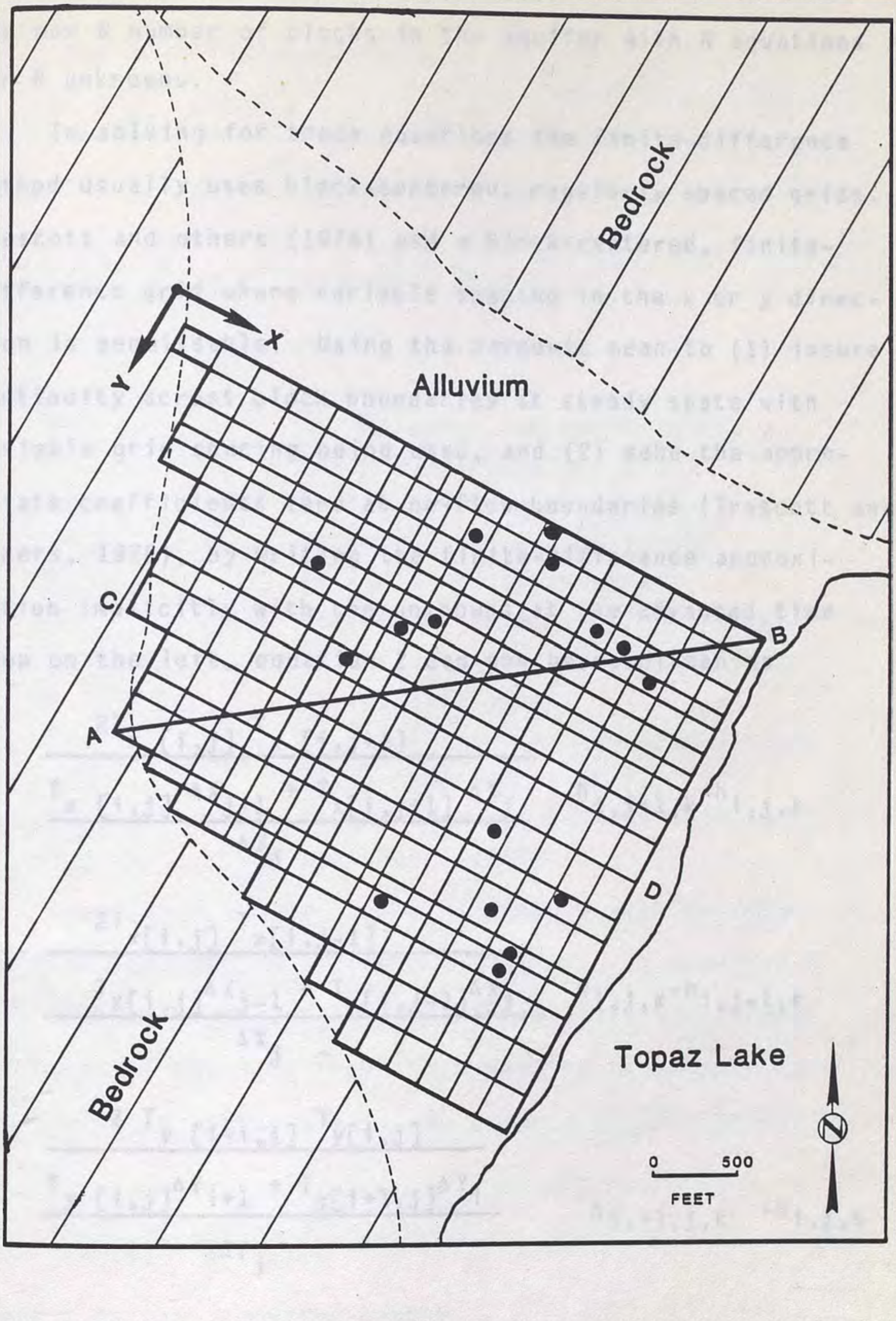


Figure 20. Computer Model Grid and Cross-Sections A-B and C-D.

are now N number of blocks in the aquifer with N equations for N unknowns.

In solving for these equations the finite-difference method usually uses block-centered, regularly spaced grids. Trescott and others (1976) use a block-centered, finite-difference grid where variable spacing in the x or y direction is permissable. Using the harmonic mean to (1) insure continuity across block boundaries at steady state with variable grid spacing being used, and (2) make the appropriate coefficients zero at no-flow boundaries (Trescott and others, 1976), by writing the finite-difference approximation implicitly with the unknowns at the advanced time step on the left, equation 2 can now be rewritten as

$$\frac{\dfrac{2T_{x[i,j]}\,T_{x[i,j+1]}}{T_{x[i,j]}\Delta X_{j+1} + T_{x[i,j+1]}\Delta X_j}}{\Delta X_j} \quad h_{i,j+1,k} - h_{i,j,k}$$

$$- \frac{\dfrac{2T_{x[i,j]}\,T_{x[i,j-1]}}{T_{x[i,j]}\Delta X_{j-1} + T_{x[i,j-1]}\Delta X_j}}{\Delta X_j} \quad h_{i,j,k} - h_{i,j-1,k}$$

$$+ \frac{\dfrac{2\,T_{y[i+i,j]}\,T_{y[i,j]}}{T_{y[i,j]}\Delta Y_{i+1} + T_{y[i+1,j]}\Delta Y_i}}{\Delta Y_i} \quad h_{i,+1,j,k} \quad - h_{i,j,k}$$

$$- \frac{\dfrac{2\, T_{y[i,j]}\, T_{y[i-1,j]}}{T_{y[i,j]}\, \Delta Y_{i-1} + T_{y[i-1,j]}\, \Delta Y_{i}}}{\Delta Y_{i}} \left(h_{i,j,k} - h_{i-1,j,k} \right)$$

$$= \frac{S_{i,j}}{\Delta t} \left(h_{i,j,k} - h_{i,j,k-1} \right) + W_{i,j,k} \qquad (4)$$

where

Δx_{j}	is the space increment in the x direction for column j (L);
Δy_{i}	is the space increment in the y direction for column i (L);
t	is the time increment (t);
i	is the index in the y direction;
j	is the index in the y direction;
k	is the time index;
$T_{x(i,j+1/2)}$	is the transmissivity between node (i,j) and node (i,j+1);
$x_{j+1/2}$	is the distance between node (i,j) and node (i,j+1).

Equation 4 is the finite difference approximation for the flow equation in a confined aquifer. In equation 4, if S is changed to S_y and the transmissivities are defined as functions of the head from the preceding iteration, equation 3 can also be approximated. An example of this replacement is

$$T^{n}_{x(i,j)} = K_{x}\, b^{n-1}_{i,j,k} \qquad (5)$$

where n is the iteration number.

AQUIFER CHARACTERISTICS FOR THE MODELED SYSTEM

In computer modeling of hydraulic systems, the flow of water is usually considered to be either lateral or vertical. This flow is termed flux and represents the flow of water through a given cross-sectional area in a designated time period. If a total difference between flux into a system (recharge), and total flux out of a system (discharge) exists, the result is a change in the water levels within the aquifer inducing changes in storage. The phenomena which could induce such a change influence flux conditions and are referred to as hydraulic stresses.

When an aquifer is subjected to pumping or recharge, its response will depend upon the geologic and hydrologic conditions, aquifer properties, and the extent of pumping or recharge. For a model to reproduce historical data correctly, all of the above parameters must be defined and entered into the model.

The specific items needed in the model include boundary conditions, hydraulic conductivities, saturated thickness, and intitial head distributions throughout the system. Because of limitations in availability and accuracy of basic data, some compromises and assumptions were made in the data input. The following sections describe the origin and formulation of the input data to the model.

Boundary Conditions

Boundary conditions encountered in flow thorugh porous media usually deal with either the potential along a boundary (Dirichlet condition) or the derivative of the potential (Neumann condition) as a flux rate specified along a boundary.

The Dirichlet condition exists where the head or potential is known as either a constant or a variable along a boundary where the aquifer is in direct hydraulic contact with a lake or river. For further explanation of how the model handles this boundary, see user documentation in Trescott, and others (1976). At Topaz Lake this boundary is found along the shoreline, but instead of a prescribed potential boundary, a fluctuating head is encountered responding to changes in lake stage.

To account for a variable prescribed potential boundary, some changes in the program logic were needed. This was accomplished by adding several statements in the subroutine which initializes data for the subsequent time step. These statements replace the prescribed potential values from the previous time step with values for the next one. By doing this, the gradient reversals seen in the historic data can be simulated, resulting in seepage into and out of the aquifer depending upon the lake stage. These changes are discussed in the Flux Term section.

The Neumann condition is employed at each node along a boundary where the flux normal to the boundary surface is specified. This flux may be either a constant equal to zero or a finite value. If it is equal to zero, then it represents a no flow or impermeable boundary. When the flux is given a finite value, the boundary becomes a constant flux, or recharge boundary. This type of boundary is treated in the model by using the corresponding nodes as sites of recharge wells with some constant recharge rate.

The western boundary of the study area where both the consolidated and unconsolidated rocks meet, is where constant recharge was employed. Here, various recharge rates were used in a trial and error fashion until a satisfactory simulation of historic data was achieved.

These rates were varied from that estimated by the Maxey-Eakin method, 144 acre-feet per year, to ones an order of magnitude greater. When 144 acre-feet/year was used, the resulting heads were about five feet lower then the historic data on the upper portions of the fan. The rates were then increased until satisfactory reproduction of historic data was produced. This value was found to be 1000 acre-feet per year, a value about 7 times greater than predicted by the Maxey-Eakin method.

This underestimation of the recharge rate to the aquifer is a reasonable difference considering the origin of the method. It was implemented and formulated in southern

Nevada to be used for large drainage basins, not small ones within a large basin. The basis of the method does not take into consideration geologic conditions (surface or subsurface), so being within an order of magnitude is an acceptable error.

Hydraulic Conductivity

To estimate hydraulic conductivity, the plots of well heads versus lake stage were used to determine the phase lag (Appendix 1). After evaluating the lag time for each well, the distance to the lake was then measured and a value for hydraulic conductivity was calculated.

The values obtained by this method were used as initial estimates during early stages of model development. Ultimately, however, these had to be adjusted, as part of the calibration process in order to simulate the historic data collected. As a starting point, an average value of 60 feet per day was used throughout the aquifer.

These were then varied within real world limits and compared to adjacent valleys. Rush and Schroer (1975) reported average values of 60 feet per day in alluvial fan material of Smith Valley to the north. In using this comparison and simulations employing varying values, the average conductivity used for the model was chosen as 52 feet per day. This value for conductivity used is nearly equal to the rate of travel of an induced pulse in the aquifer, 60 feet day, determined with the aid of the phase lag.

This lower value enabled reproduction of the water table including a pulse sent through the aquifer induced from the fluctuating lake levels. This is shown in Figure 21 when the rising lake had induced a pulse which was

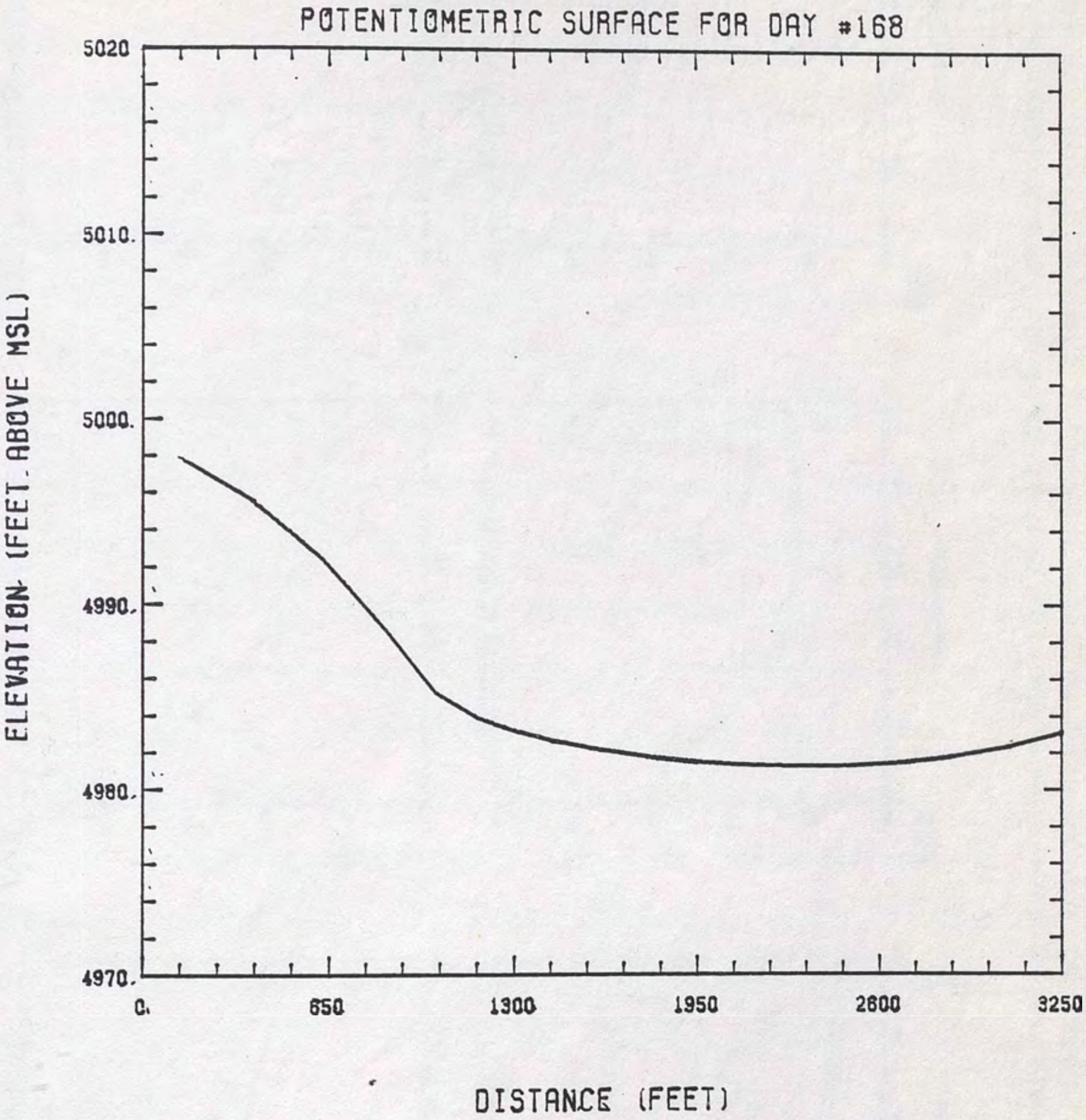


Figure 21. Piezometric Surface Cross-Section C-D.

reproduced by the model. Figure 22 depicts a time series of cross-section C-D, Figure 20, illustrating the change in piezometric surface with time as the lake stage fluctuates.

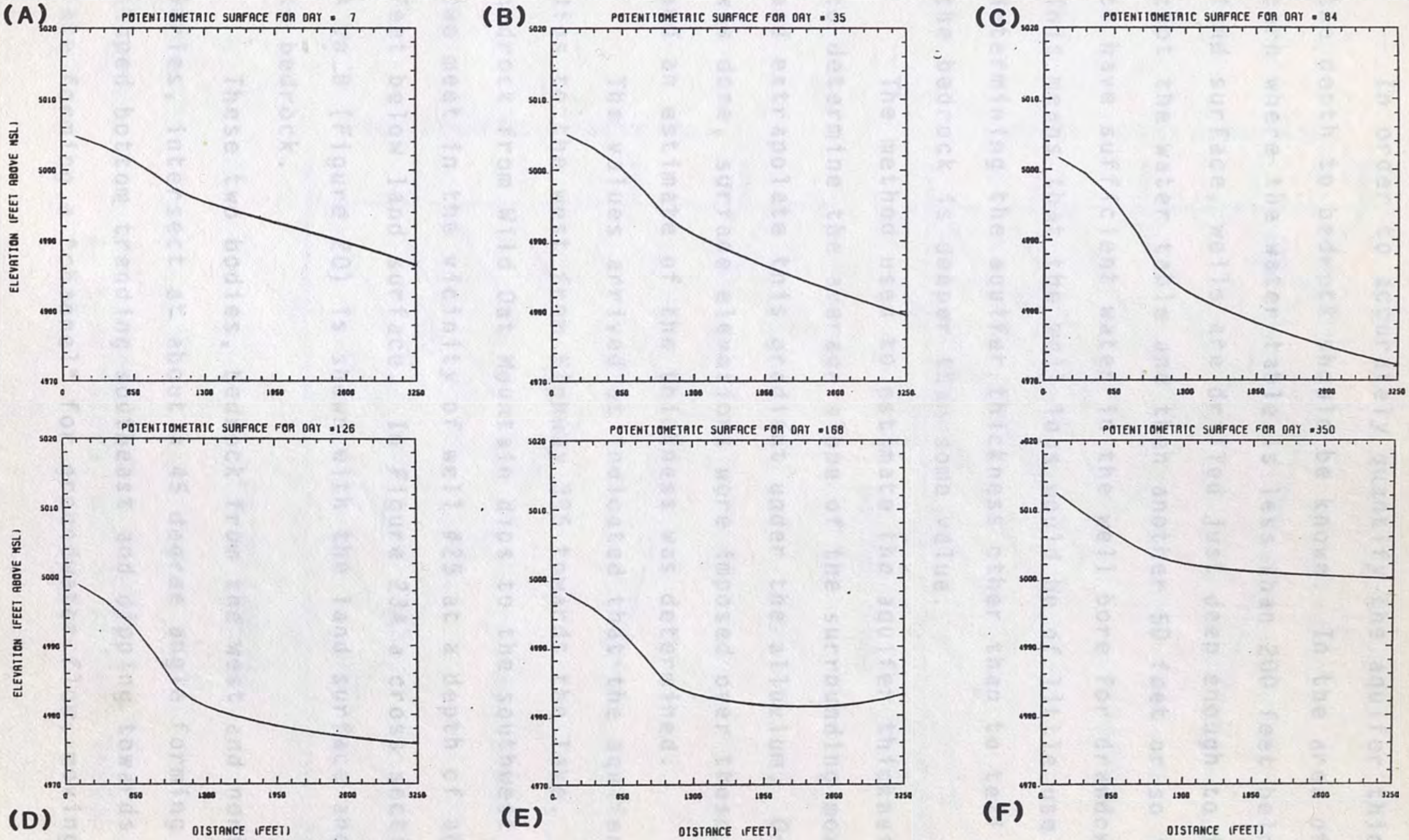


Figure 22. Piezometric Surface Cross-Section C-D at Various Lake Stages; A) Declining Lake Stage, B) Approaching Minimum Stage, C) Minimum Stage, D) Lake Starts to Rise, E) Gradient Reversal is Seen, F) Maximum Stage.

Saturated Thickness

In order to accurately quantify the aquifer thickness, the depth to bedrock should be known. In the area of concern where the water table is less than 200 feet below the land surface, wells are drilled just deep enough to intercept the water table and then another 50 feet or so in order to have sufficient water in the well bore for drawdown. This means that the well logs would be of little use in determining the aquifer thickness other than to tell that the bedrock is deeper than some value.

The method used to estimate the aquifer thickness was to determine the average slope of the surrounding mountains and extrapolate this gradient under the alluvium. Once this was done, surface elevations were imposed over these values and an estimate of the thickness was determined.

The values arrived at indicated that the aquifer bottom dips to the west from Highway 395 towards the lake. The bedrock from Wild Oat Mountain dips to the southwest and the two meet in the vicinity of well #25 at a depth of about 550 feet below land surface. In Figure 23A a cross section from A to B (Figure 20) is shown with the land surface and depth to bedrock.

These two bodies, bedrock from the west and north boundaries, intersect at about a 45 degree angle forming a V-shaped bottom trending southeast and dipping towards the lake forming a "channel" for groundwater flow, moving

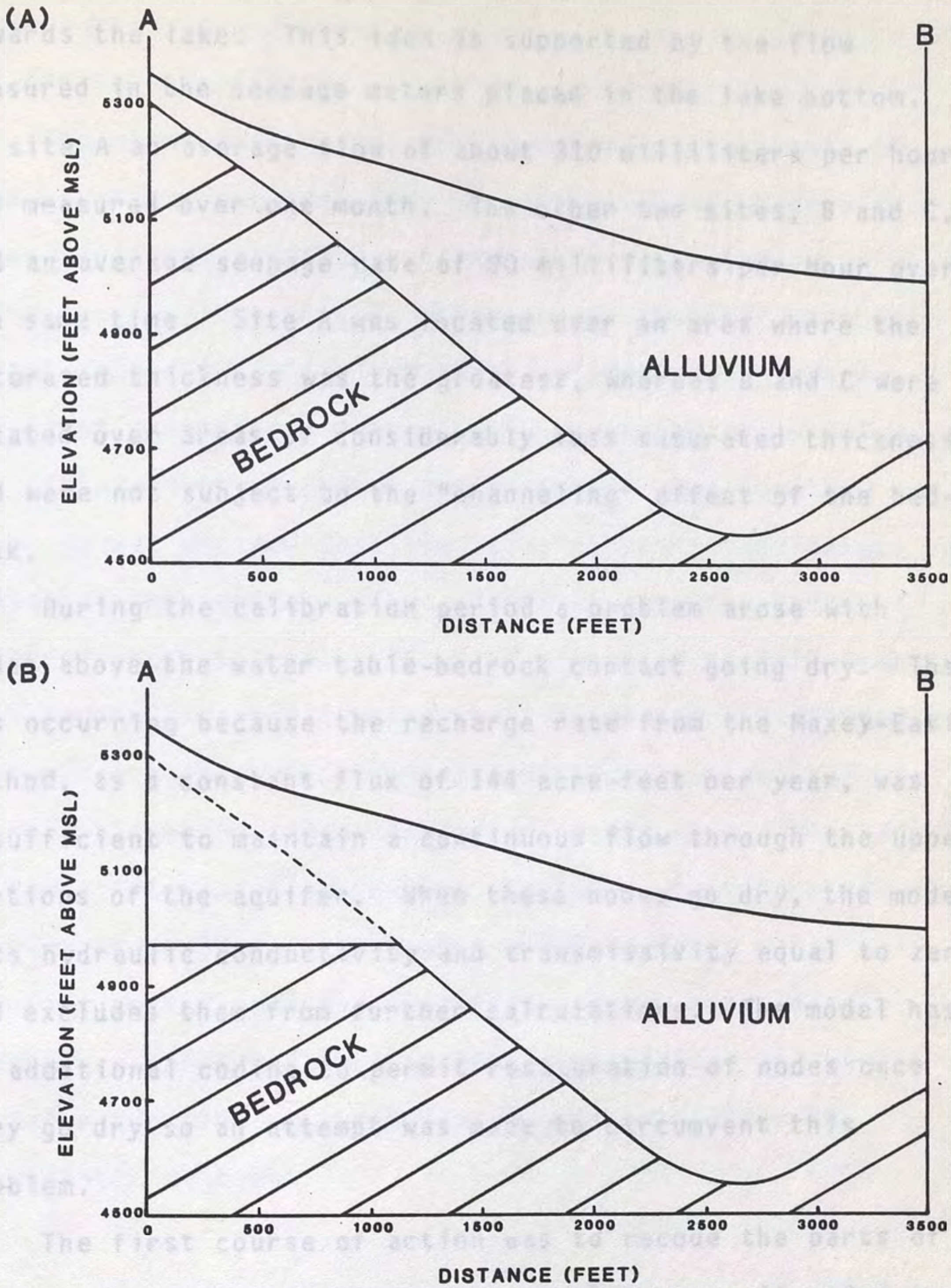


Figure 23. Cross-Section A-B of Alluvium Showing, A) True Values, and B) Assumed Values for Model.

towards the lake. This idea is supported by the flow measured in the seepage meters placed in the lake bottom. At site A an average flow of about 310 milliliters per hour was measured over one month. The other two sites, B and C, had an average seepage rate of 20 milliliters per hour over the same time. Site A was located over an area where the saturated thickness was the greatest, whereas B and C were located over areas of considerably less saturated thickness and were not subject to the "channeling" effect of the bedrock.

During the calibration period a problem arose with nodes above the water table-bedrock contact going dry. Ths was occurring because the recharge rate from the Maxey-Eakin method, as a constant flux of 144 acre-feet per year, was insufficient to maintain a continuous flow through the upper portions of the aquifer. When these nodes go dry, the model sets hydraulic conductivity and transmissivity equal to zero and excludes them from further calculations. The model has no additional coding to permit resaturation of nodes once they go dry so an attempt was made to circumvent this problem.

The first course of action was to recode the parts of the program which set parameters equal to zero if nodes went dry. This was attempted but resulted in large descrepancies in head values and inconsistencies in mass balances. It was decided that in order to recode the program to allow for resaturation of nodes, it would require rewriting the logic for the entire program.

The other alternative was to manipulate the type of flow boundary, constant flux or constant head, hydraulic conductivities, and flux rate for the constant flux boundary. Various scenarios were used, incorporating different combinations of increments to each of the aforementioned parameters, none of which resolved the problem.

In order to reproduce the historic potentiometric surface, it was decided that the solution was to manipulate the bottom of the aquifer. The assumption that the bedrock-alluvium contact levels off at an elevation of 4970 feet is depicted on Figure 23B where the true bottom and assumed bottom are both shown. This assumption was made so that there would be no nodes that could go dry when a constant flux was introduced on the west boundary of study area. This was necessary so that when the lake level rises inducing an increase in water table level, the nodes farthest up on the fan could remain saturated, allowing recharge from the constant flux boundary to enter the groundwater system.

Head Distribution

The head values needed to determine saturated thickness and transmissivity were compiled from water levels in the wells. This collection was done for a period of one year, normally on a weekly basis, and represents a good historical data base to use in model calibration (Appendix I).

In comparing collected lake levels and heads in wells along the shoreline, a five foot discrepancy was found to exist with the lake always being five feet higher. In a subsequent field reconnaissance results indicated a difference was present in the data collected but not actually in the field.

From this it was deduced that either the highway marker from which the survey was done, or the datum point where the lake levels were read, was five feet off.

Conceptually in the model it would make no difference whether the lake levels were lowered five feet or the well levels raised five feet, so, as a matter of convenience, the lake levels were all lowered by five feet.

The starting head matrix used in the final model simulations was obtained from a previous simulation. Lake heads for the prior year were used for the constant head boundary at the lake and the model was run for a period of 304 days ending with the first day of the study period.

This final head matrix was then used in a new data set as the starting head matrix for the study period. In doing this, the initial heads used were already in phase with the fluctuating head boundary and the systems mass balance would not have to equilibrate itself.

FLUX TERM

One of the major problems with the model was to allow a changing constant head boundary along the shoreline of the lake. To do this, modifications were made to the program logic to reset the values for head at the nodes along the shore to reflect fluctuations in lake stage.

After each time step, a new lake level was read in and PHI(I,J), for the nodes involved was assigned the new value. One result of this change was that the mass balance computations for flow to and from a constant head boundary were adversely affected. When these computations are done, the head difference is calculated as the starting head at the constant head node minus the head of the adjacent node at the same time step, STRT (I,J)-PHI(I,J-1).

Since the starting head value does not get reassigned, the calculations were indicating a large gradient towards the lake, when recharge was actually occurring. This calculation was changed so that the head from the adjacent node for the previous time step, KEEP (I,J-1), was subtracted from the new lake level read in for the current time step, giving the equation as PHI(I,J)-KEEP)(I,J-1).

This change allowed for the lake body, acting as a constant head boundary, to have flux into and out of the aquifer when the lake was rising or declining.

OBSERVED VERSUS SIMULATED CONDITIONS

Prior to this model being used to predict future response of the aquifer to imposed stresses from changing lake stage, it was necessary that it be able to reasonably duplicate observed response to historical stresses. This evaluation was done by continually comparing simulated water levels and water budget output to field measurements and observations.

Data prior to this study were collected intermittently, usually only several times per year. Because of this, there was insufficient historical data on which to assess the validity of the transient model other than the collected data during the study period. Data on recharge to the aquifer itself were virtually nonexistent and presented problems to quantifying and validating the estimates for recharge used in the model. As a result of the lack of reliable data, the modeled quantities used can only be assessed for their reasonableness in reproducing observed water-level conditions in the area.

The accuracy of the transient simulation was checked by a comparison of simulated water-levels with those measured at the same point in time and space. It was found that the model results were unable to reproduce the observed heads at wells #14 and #15. During the calibration period, the values for the aquifer characteristics were varied in an attempt to obtain the observed heads but to no avail. It was then decided that these were anomalous zones under influence from other unknown hydrologic and/or geologic constraints. Those zones are located farthest from the shoreline and closest to the fault defined bedrock-aquifer contact. This faulted zone could have local fractures into which the hidden recharge discussed earlier could be channeled to well #15 or away from well #14.

Aside from wells #14 and #15, the result of the comparison was that the modeled heads closely reproduced the historical data and that the differences were small enough so that any further attempt to improve the model regarding this error was deemed impractical and unjustified (Figure 24). The errors present are theoretically an indication of the limited data with which the model was calibrated in the steady-state solution and are well within acceptable limits for model errors.

On Figure 25, a comparison of historical and simulated contour maps was used to show the changing water table configuration as the reservoir stage fluctuated. From these, the contour lines are seen to either shift upgradient as the reservoir was rising or downgradient as it was declining. The historic and simulated maps are not identical because of the linear interpolation in the computer plotting routine and the fact that the location of data points for the simulated plots were calculated by the computer model. These data points were block-centered and did not represent the

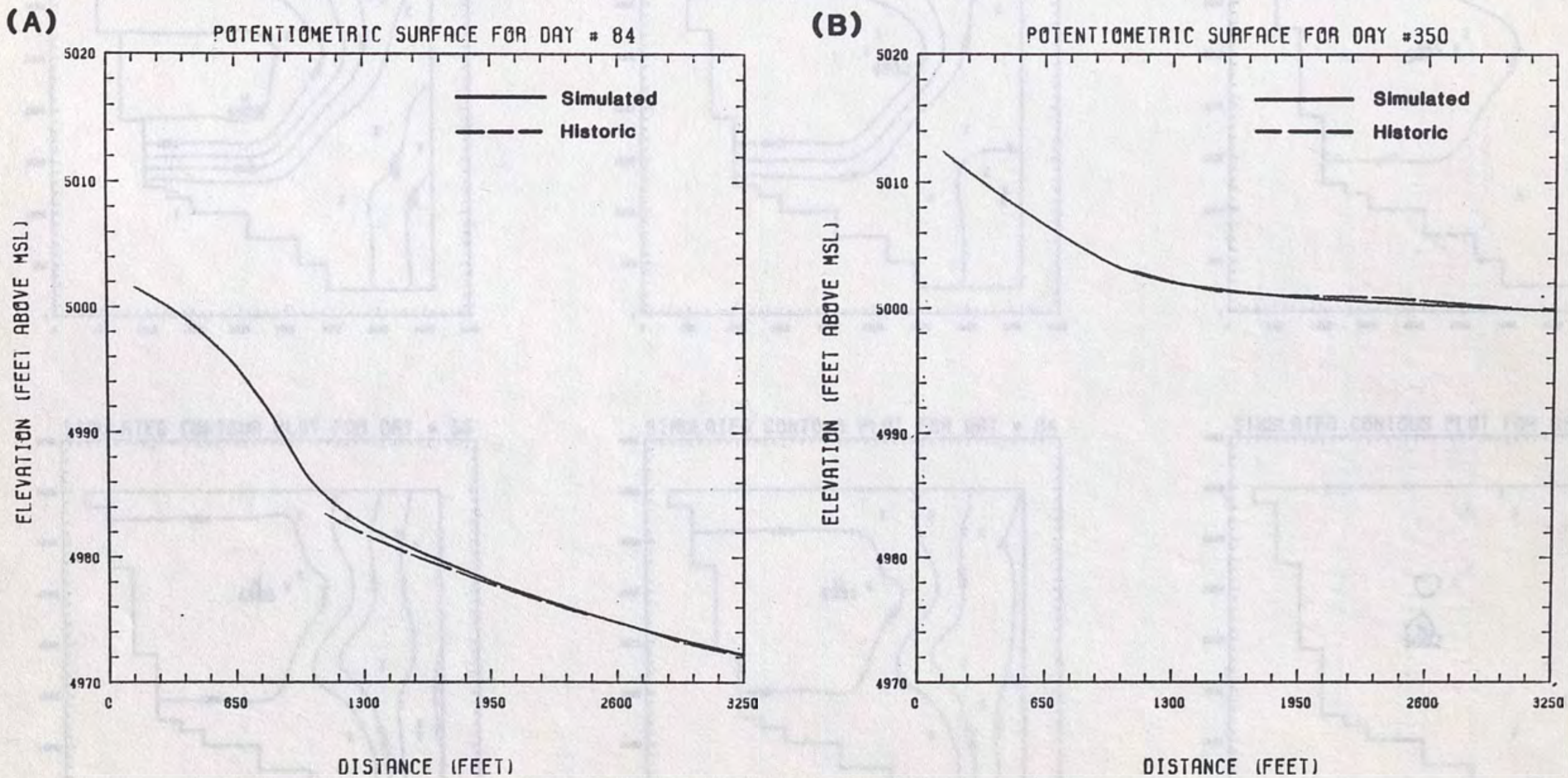


Figure 24. Cross-Section C-D Comparison of Historic and Simulated Head Distributions for A) Day #84, and B) Day #350.

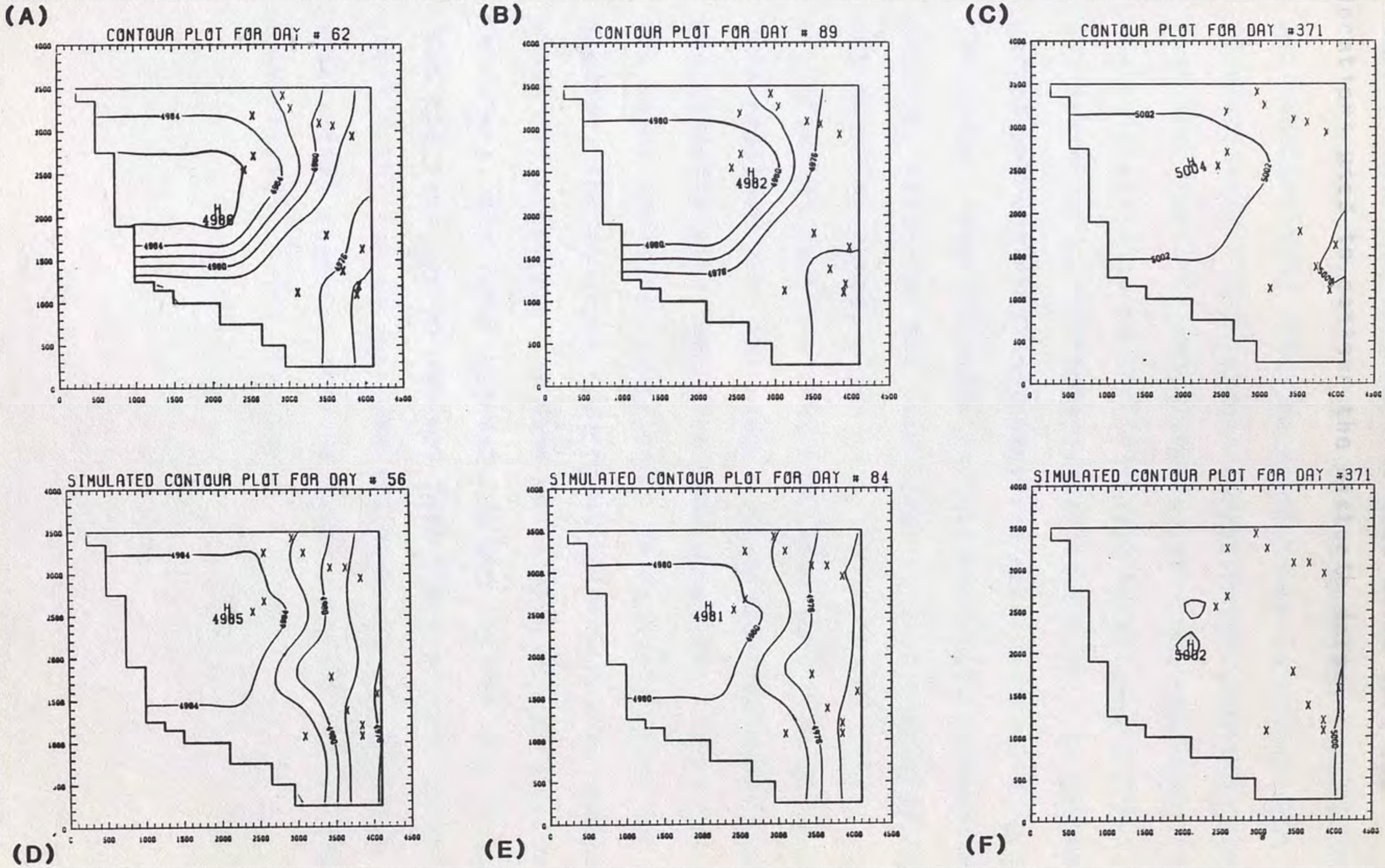


Figure 25. Comparison of Historic and Simulated Contour Maps of the Piezometric Surface for Selected Days.

true well locations within that node which were the locations used to contour the historic data.

MASS BALANCE ANALYSIS FOR RECHARGE

A cumulative mass balance dealing with constant flux and constant head into the aquifer was calculated during simulations. These values represent the amount of recharge from the fractured consolidated rocks and the lake. This analysis pertains to the 1979-1980 water year and reflects the hydrologic conditions during this time. It covers a period where at the beginning the reservoir was at a fairly low stage, then increased to full capacity, followed by a decline, returning the lake level to within several feet of its starting level.

For the constant flux boundary, an influx of about 865 acre-feet per year was used. It was divided among nodes at the boundary with some receiving more than others so that the heads could be reproduced. To maintain heads in the aquifer, the constant head boundary at the lake contributed about 356 acre-feet per year to the aquifer as a net inflow. Therefore, the total recharge to the aquifer was 30 percent from the lake and 70 percent from the fractured consolidated rocks. The percentage from the lake represents a substantial quantity of lake water entering the aquifer as bank storage during the year.

MASS BALANCE ANALYSIS FOR DISCHARGE

A cumulative mass balance was also calculated for discharge from the aquifer as flux into the reservoir from the aquifer. It was implemented for the 1979-1980 water year and reflects hydrologic conditions during this time.

During this water year, a net outflux from the aquifer to the reservoir of 920 acre-feet was calculated to have been discharged. This value is approximately 300 acre-feet less than the recharge amount to the aquifer indicating that for this water year a net increase in aquifer storage of 300 acre-feet occurred. This increase in storage is reflected in the water table being about 7 feet higher than it was the previous year.

SOLUTE TRANSPORT ANALYSIS

A computer model for evaluating two-dimensional solute transport and dispersion in groundwater by Konikow and Bredehoeft (1978), was planned to be used in this study. Insufficient time for modification of program logic and model calibration prevented its use, so a simple analysis of a qualitative conceptual model and data for the area was done. It was used to analyze the solute transport and dispersion of a contaminant, such as nitrate, in an assumed initial concentration of 100 mg/l. This value was chosen as one greater than the maximum allowable concentration of 45 ppm and one easy to work with.

The volume of water contributed to the aquifer from both consolidated rocks and the lake, and the amount of outflux from the aquifer into the reservoir, were calculated by the two-dimensional flow model. Changes in lake storage were interpolated from Figure 11 and were used to estimate the extent of dilution or enrichment of the contaminant.

The first scenario will consider the possibility of the groundwater degrading reservoir quality as the aquifer is being drained of water with an initial concentration of 100 ppm by a declining lake stage. During this portion of the cyclic seasonal fluctuations, the lake declined 28 feet, resulting in a flux of about 500 acre-feet into the reservoir.

For a conservative evaluation, it will be assumed that groundwater input only mixed with the 36000 acre-feet released from the reservoir during this time. The resultant reduction in concentration from a mixing ratio of 1:72, was from 100 to 1.4 ppm, well below the maximum allowable concentration.

With this mixing ratio, the contaminant concentration in the aquifer would have to be extremely high before any appreciable degradation of the reservoir would be seen. The result of this analysis is that the potential for reservoir degradation from the aquifer becoming contaminated is unlikely because of the mixing ratio and the short residence time in such a flow-through reservoir.

The next scenario of interest would be to consider the potential for aquifer degradation from the reservoir becoming contaminated. Two analyses will be presented; one with flow into the aquifer as piston flow, advection, of a slug of reservoir water equal to 285 acre-feet and the others as flow influenced by dispersion. Dispersion is the irreversible phenomenon where the solute transport regime extends beyond the flow domain defined by piston flow.

With advection, the slug would migrate into the aquifer during the bank storage period as a separate entity and then be flushed out during the next decline cycle. If the reservoir contamination was a one time occurrence while bank storage was in progress, the following bank storage cycle would contribute only clean water to the aquifer since the contaminant would have been flushed out of the reservoir. If dispersion was introduced, it would require more time, but the contaminant would eventually be flushed out of the aquifer, some with each lake decline cycle. Figure 26 shows a qualitative conceptual picture of this occurring in an idealized model, one where no actual numbers can be assigned.

Should the reservoir become contaminated continuously from an upstream source, piston flow would move a slug into the aquifer with each cycle of bank storage, and flush it out with each decline. With dispersion, this scenario would create an undesirable condition. As the reservoir infiltrated the aquifer, the extent of degradation would be beyond the domain defined by considering only piston flow. When the reservoir stage declined, some of the solute would be removed, but because of dispersion, not all of it. With each cyclic fluctuation, the extent of aquifer degradation would increase as dispersion moved the solute farther into the aquifer. Figure 27 shows this occurring, again as an idealized conceptual model where no numbers can be assigned.

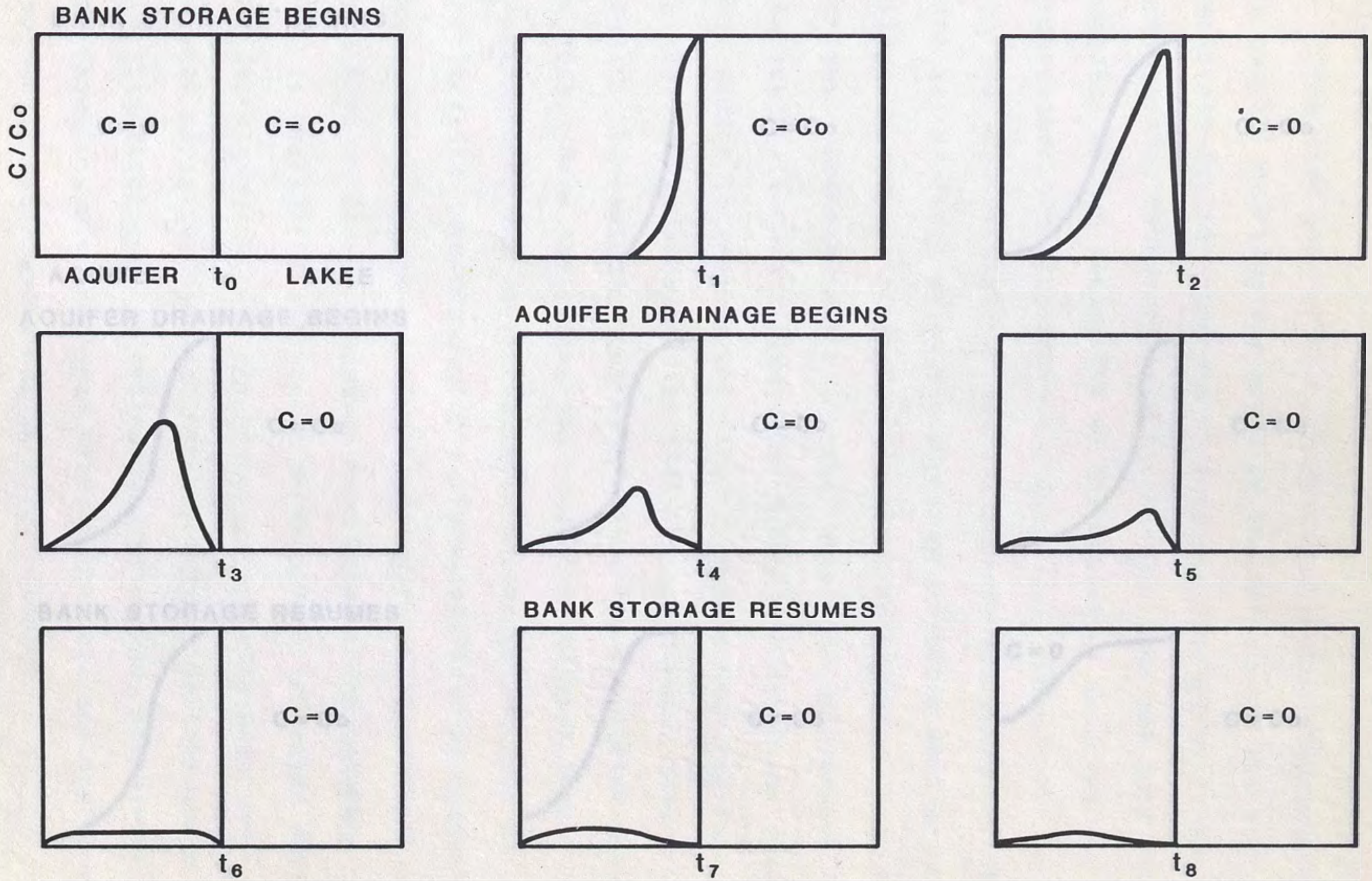


Figure 26. Qualitative Conceptual Picture of a One-Time Introduction of a Slug of Contaminant into the Reservoir with Dispersion.

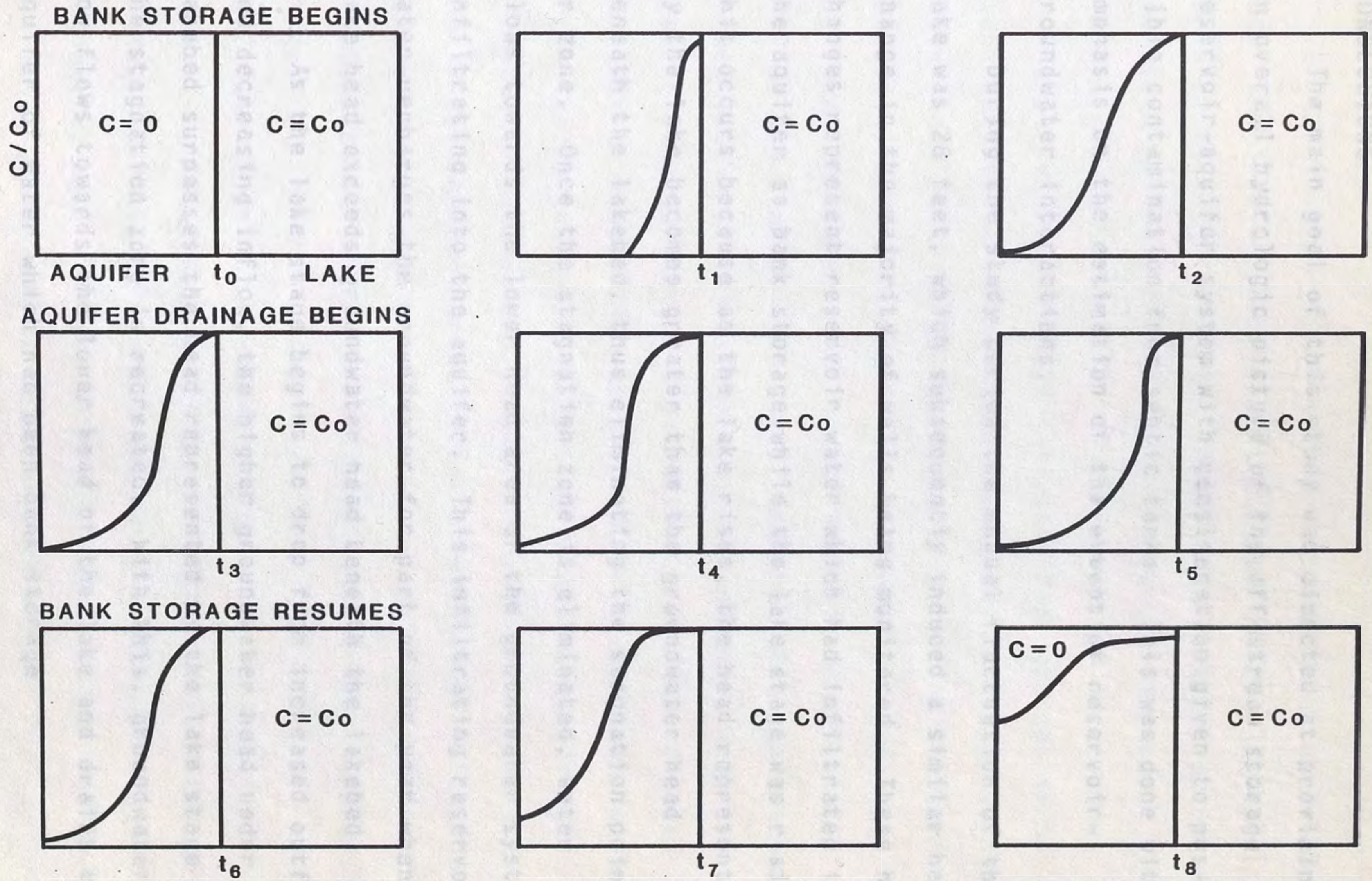


Figure 27. Qualitative Conceptual Picture of a Constant Source of Contamination to the Reservoir with Dispersion.

CONCLUSIONS

The main goal of this study was directed at providing an overall hydrologic picture of the off-stream storage reservoir-aquifer system with consideration given to possible contamination from septic tanks. This was done with emphasis on the estimation of the extent of reservoir-groundwater interactions.

During the study period the annual fluctuation of the lake was 28 feet, which subsequently induced a similar head change in the majority of wells being monitored. These head changes represent reservoir water which had infiltrated into the aquifer as bank storage while the lake stage was rising. This occurs because as the lake rises, the head represented by the lake becomes greater than the groundwater head beneath the lakebed, thus eliminating the stagnation point or zone. Once the stagnation zone is eliminated, water flows towards the lower head area of the groundwater system, infiltrating into the aquifer. This infiltrating reservoir water recharges the groundwater for part of the year when lake head exceeds groundwater head beneath the lakebed.

As the lake stage begins to drop from increased outflow and decreasing inflow, the higher groundwater head under the lakebed surpasses the head represented by the lake stage and the stagnation zone is recreated. With this, groundwater now flows towards the lower head of the lake and drains the aquifer of water which had been bank storage from the lake.

Some recharge of the aquifer results from the influx of water from the surrounding consolidated rocks. This water moves downgradient from the western and northern boundaries and mixes with the chemically different infiltrated reservoir water.

Areas which exhibited influence from changes in lake stage were the areas to which migration and mixing occurred. Above these points, the water's origin is from the surrounding fracture consolidated rocks and shows seasonal influence from spring and summer snowmelt.

At the time of this study, analyses of groundwater showed little if any evidence of contamination from the present distribution of septic tanks. With anticipated increase in residential development of the area, this condition could change and result in contamination of the groundwater from the increased number of septic tanks. Should development increase closer to the shoreline, the probability of aquifer contamination would be greater because of the shallower water table at periods of high lake stage.

Results from the computer simulation show that 30 percent of total aquifer recharge is contributed from the rising prescribed potential boundary. This quantity of influx, as bank storage, is sufficient to result in the migration and mixing of this water throughout most of the aquifer. As the lake level declined, most of this water would be drained from the aquifer back into the lake. Then as the reservoir began to rise again, recharge to the aquifer would be reinstated. As the lake prescribed potential boundary declined, 75 percent of the water recharged to the aquifer was discharged resulting in the water table being at a higher level than the previous year.

Considering the existence of such a good hyraulic connection between the two bodies of water, a real problem would present itself should the reservoir become contaminated. If it was to occur during the time when bank storage was being introduced to the aquifer, groundwater used from the domestic wells in the area would in all probability become contaminated.

The extent of this contamination would not be confined to the flow domain of the recharged water, rather it would also be influenced by the effects of dispersion. This spreading phenomenon where a solute spreads out with time, would result in the solute occupying an ever increasing volume of the flow domain, one larger than that defined by just considering advection. Even if the flow caused contaminant migration into a small portion of the aquifer, dispersion would cause the degradation of an additional volume uf the flow domain.

REFERENCES

Bredehoeft, J.D., and Pinder, G.F., 1970, Digital analysis of areal flow in multiaquifer groundwater systems: A quasi three-dimensional model: Water Resources Research, v.6, no.3, p. 883-888.

Carlson, J.E., Stewart, J.H., Johannesen, Dann, and Kleinhampl, F.J., 1978, Geologic map of the Walker Lake 1°x2° quadrangle, Nevada-California: U.S. Geol. Survey open-file report, no. 78-523.

Cooper, J., 1980, Research Associate, Bioresources Center, Desert Research Institute, Reno, NV.

Curtis, G.H., 1951, The geology of the Topaz Lake quadrangle and the eastern half of the Ebbetts Pass quadrangle: Univ. of California Ph.D. thesis.

Eakin, T.E., and others, 1951, Contributions to the hydrology of eastern Nevada: Nevada Dept. Conserv. and Nat. Resources: Water Resources Bull. 12.

Eisenlohr, W.S., and others, 1972, Hydrologic investigations of prairie potholes in North Dakota, 1959-1968: U.S. Geol. Survey Prof. Paper 585-A.

Feth, J.H., 1963, Hidden recharge: Groundwater, v.2, #4, p. 14-17.

Gilbert, C.M. and Reynolds, M.W., 1973, Character and chronology of Basin development, western margin of the Basin and Range province: Geol. of America Bull, v.84, no. 8, p. 2489-2510.

Glancy, P.A., 1971, Water-resources appraisal of Antelope Valley and east Walker area, Nevada and California: Nevada Dept. of Conservation and Natural Resources, Water Resources-Reconnaissance Series Rept. 53.

Hackbarth, D.A., 1968, Hydrogeology of the Little St. Germain Lake Basin, Vilas County, Wisconsin: Univ. of Wisconsin, Madison, unpublished M.S. thesis.

Harrill, J.R., and Nowlin, J.O., 1976, Groundwater data near the northwest shore of Topaz Lake, Douglas County, Nevada, U.S. Geol. Survey open-file report, no. 76-90.

Hennings, R.G., 1974, A hydrologic investigation of the Pickerel Lake Basin, Portage County, Wisconsin: Univ. of Wisconsin, Madison, M.S. thesis.

Houston, C.E., 1950, Consumptive use of irrigation water by crops in Nevada: Nevada Univ. Agr. Expt. Sta. Bull 185.

Kohler, M.A., Nordenson, T.J., and Baker, D.R., 1959, Evaporation maps for the United States: U.S. Weather Bur. Tech. Paper 37.

Langbein, W.B., Hains, C.H., and Culler, R.C., 1951, Hydrology of stockwater reservoirs in Arizona: U.S. Geol. Survey Circular 110.

Lee, D.R., 1977, A device for measuring seepage flux in lakes and estuaries: Limnology and Oceanography 22(1).

Meyboom, P., 1967, Mass-transfer studies to determine the groundwater regimes of permanent lakes in hummocky moraine of western Canada: Journal of Hydrology, v. 5, p. 117-142.

Meyboom, P., 1966, Unsteady groundwater flow near a willow ring in hymmocky moraine: Journal of Hydrology, v. 4, p. 38-62.

Moore, J.G., 1969, Geology and mineral deposits of Lyon, Douglas, and Ormsby Counties, Nevada: Nev. Bur. of Mines and Geology Bull. 75.

Nowlin, J.O., 1976, A brief evaluation of the chemical character of water samples collected near the northwest shore of Topaz Lake, Douglas County, Nevada: U.S. Geol. Survey open-file report, no. 76-419.

Peaceman, D.W., and Rachford, H.H., Jr., 1955, the numerical solution of parabolic and elliptic differential equations: Soc. Indust. Appl. Math. Jour., v. 3, no. 1, p. 28-41.

Pinder, G.F., and Bredehoeft, J.D., 1968, Application of a digital computer for aquifer evaluation: Water Resources Research, v. 4, no. 5, p. 1069-1093.

Possin, Boyd, 1973, Hydrogeology of Mirror and Shadow Lakes in Waupaca, Wisconsin: Univ. of Wisconsin, M.S. thesis.

Robinson, T.W., 1965, Water use studies utilizing evapotranspiration tanks, in Cohen, Philip, and others, Water Resources of the Humboldt River Valley near Winnemucca, Nevada: U.S. Geol. Survey Water-Supply Paper 1975, p. 83-91.

Rush, F.E. and Hill, V.R., 1972, Bathymetric reconnaissance of Topaz Lake, Nevada and California: Nevada Div. Water Resources, Inf. Ser. Rept. 12, 1 map.

Rush, F.E., and Schoer, C.V., 1975, Geohydrology of Smith Valley, Nevada, with special reference to the water-use period, 1953-1972: U.S. Geol. Survey open-file report 75-417.

Rush, F.E., 1968, Index of hydrographic areas: Nevada Dept. Conserv. and Nat. Resources, Water-Resources Inf. Ser. Rept. 6.

Scalf, M.R., Dunlap, W.J., and Kreissl, J.F., 1977, Environmental effects of septic tank systems: U.S. Environmental Protection Agency, Ecological Research Series, no. 600/3-77-096.

Sloan, C.E., 1972, Ground-water hydrology of prairie potholes in North Dakota: U.S. Geol. Survey Prof. Paper 585-C.

Sloan, C.E., 1970, Prairie potholes and the water table: U.S. Geol. Survey Prof. Paper 700-B, p. B227-B231.

Trescott, P.C., Pinder, G.F., and Larson, S.P., 1976, Finite-difference model for aquifer simulation in two dimensions with results of numerical experiments: U.S. Geol. Survey Techniques of Water-Resources Investigations, Book 7, Chap. C1.

Trescott, P.C., 1973, Iterative digital model for aquifer evaluation: U.S. Geol. Survey open-file report.

U.S. Environmental Protection Agency, 1979, Methods for Chemical Analysis of Water and Wastes, no. EPA-600-4-79-020.

U.S. Environmental Protection Agency, 1976, Quality Criteria for Water.

U.S. Public Health Service, 1962, Drinking water standards, 1962: U.S. Public Health Service Publication 956.

White, W.N., 1932, A method of estimation ground-water supplies based on discharge by plants and evaporation from soil: U.S. Geol. Survey Water-Supply Paper 659-A.

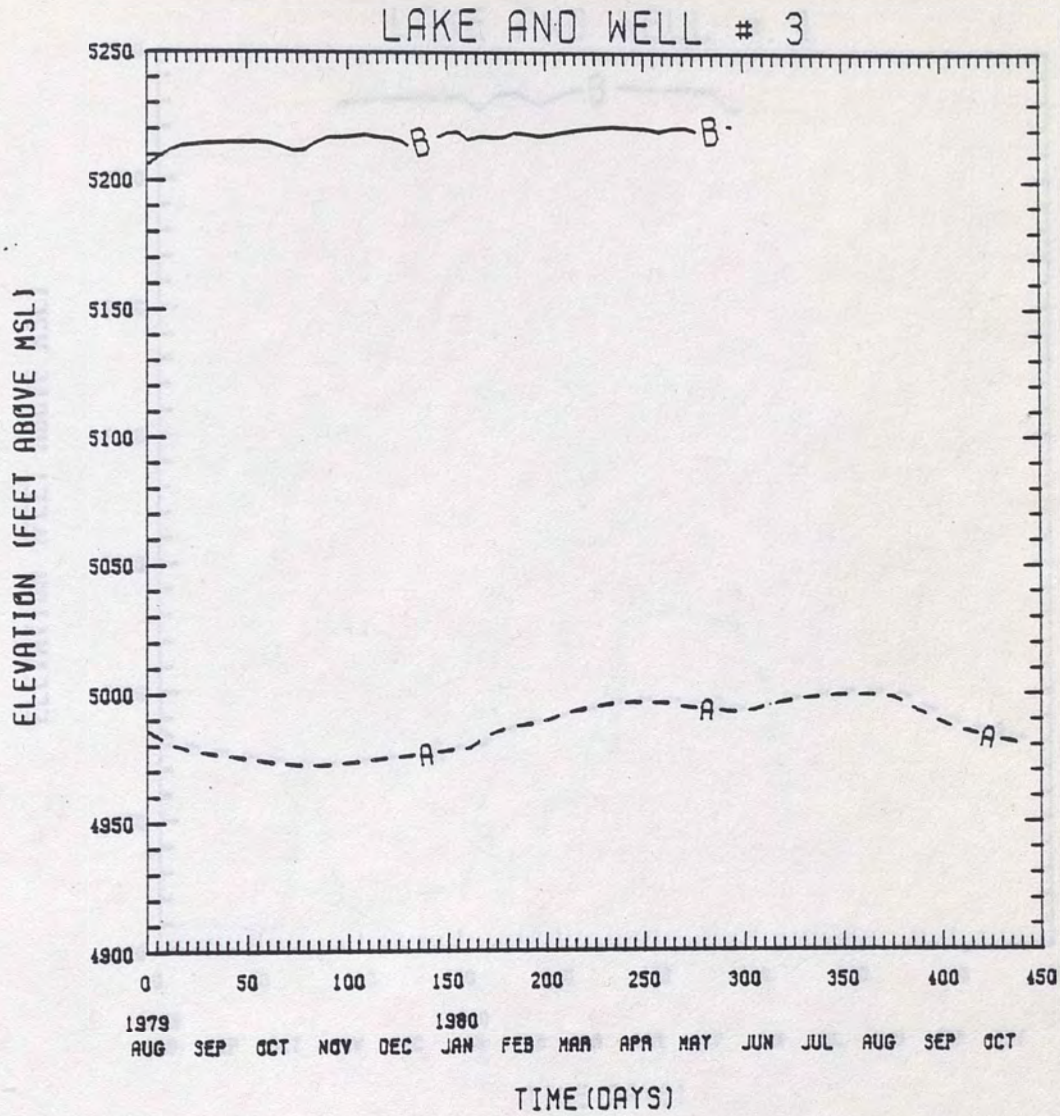
Winter, T.C., 1978, Numerical simulation of steady state three-dimensional groundwater flow near lakes: Water Resources Research, v. 14, no. 2, p. 245-254.

Winter, T.C., 1976, Numerical simulation analysis of the interactin of lakes and groundwater: U.S. Geol. Survey Prof. Paper 1001.

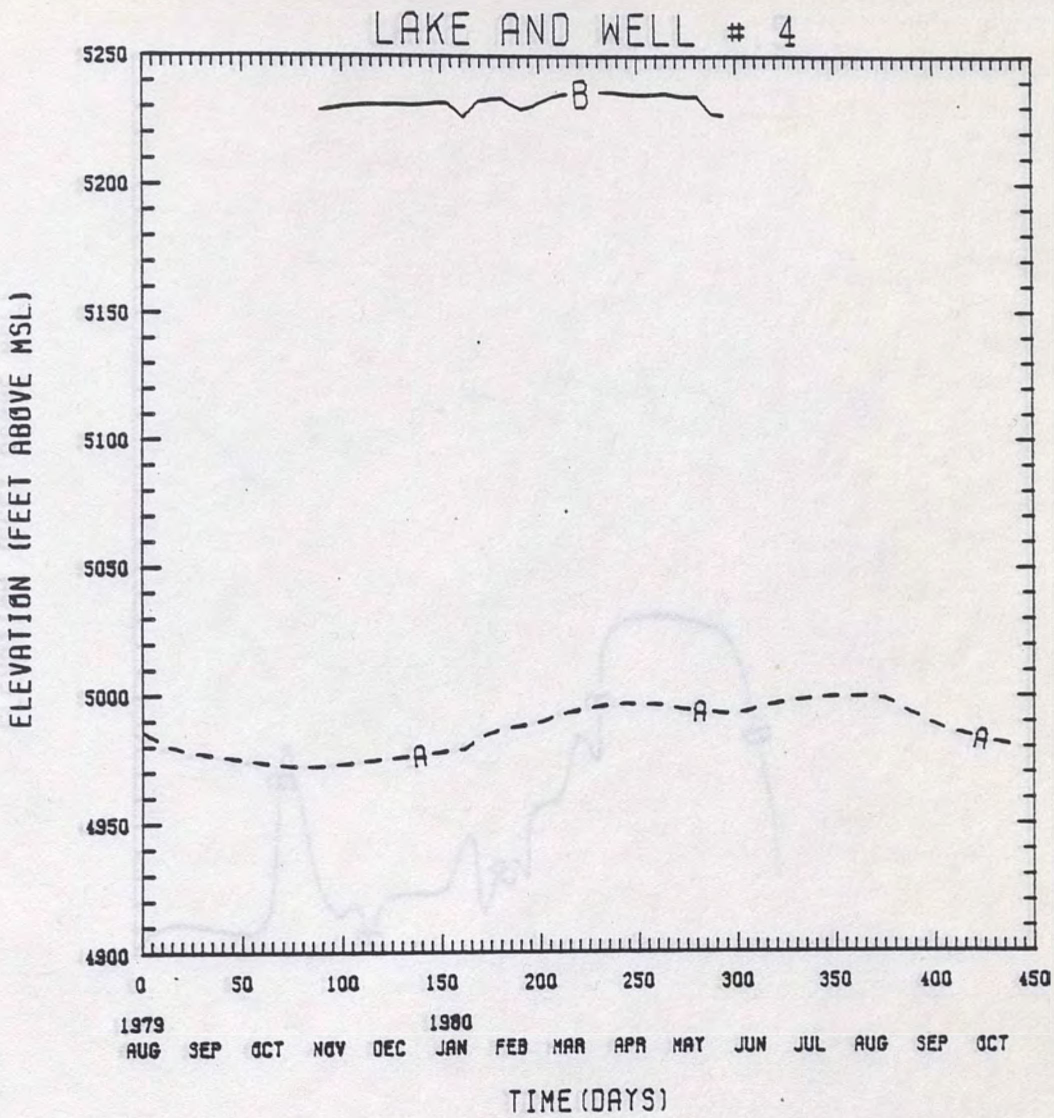
Young, A.A., and Blaney, H.F., 1942, Use of Water by native vegetation: California Div. Water Resources Bull. 50.

APPENDIX I

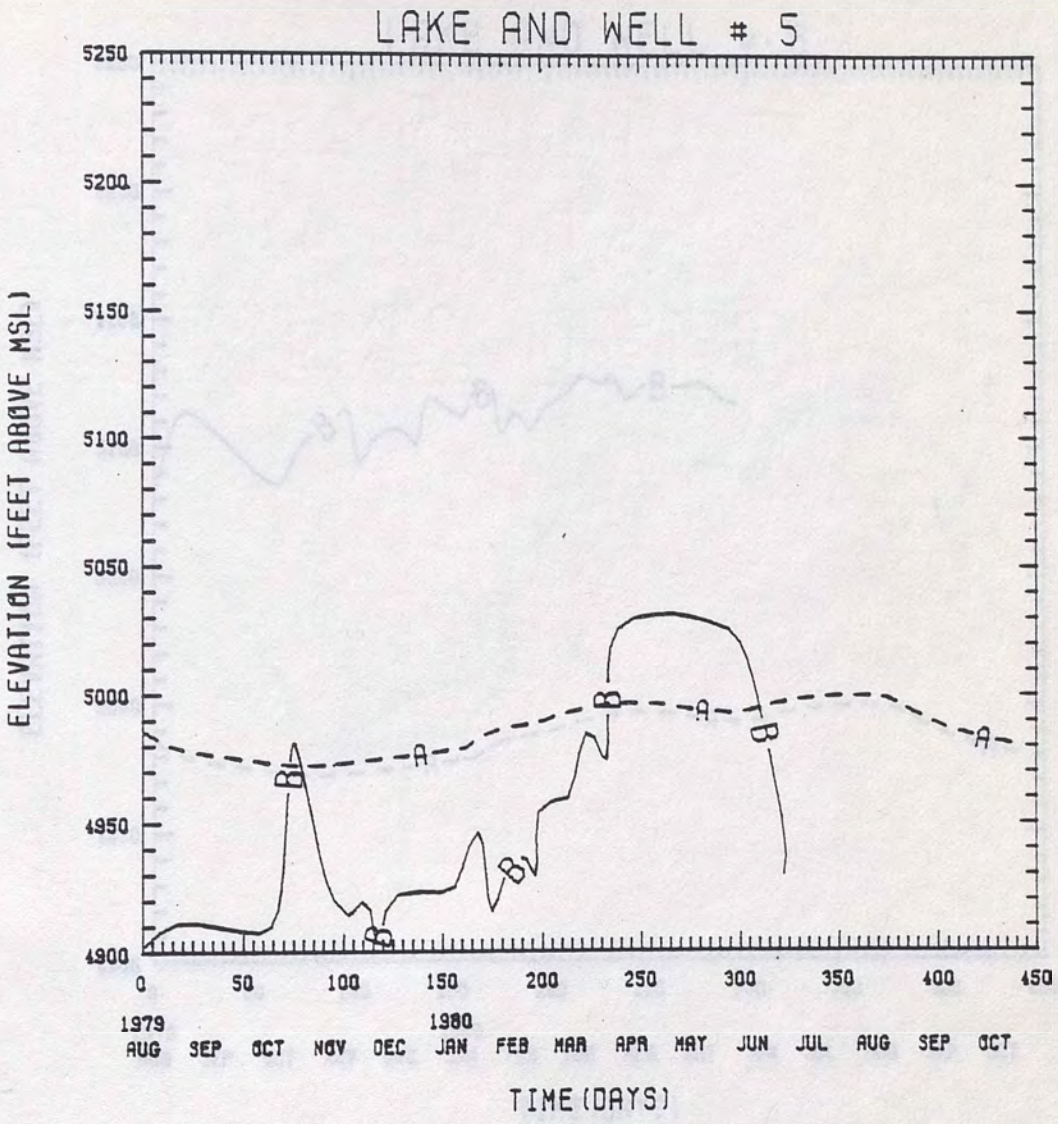
Comparison Plots of Lake Stage and Water Levels in the Monitored Wells During Study Period

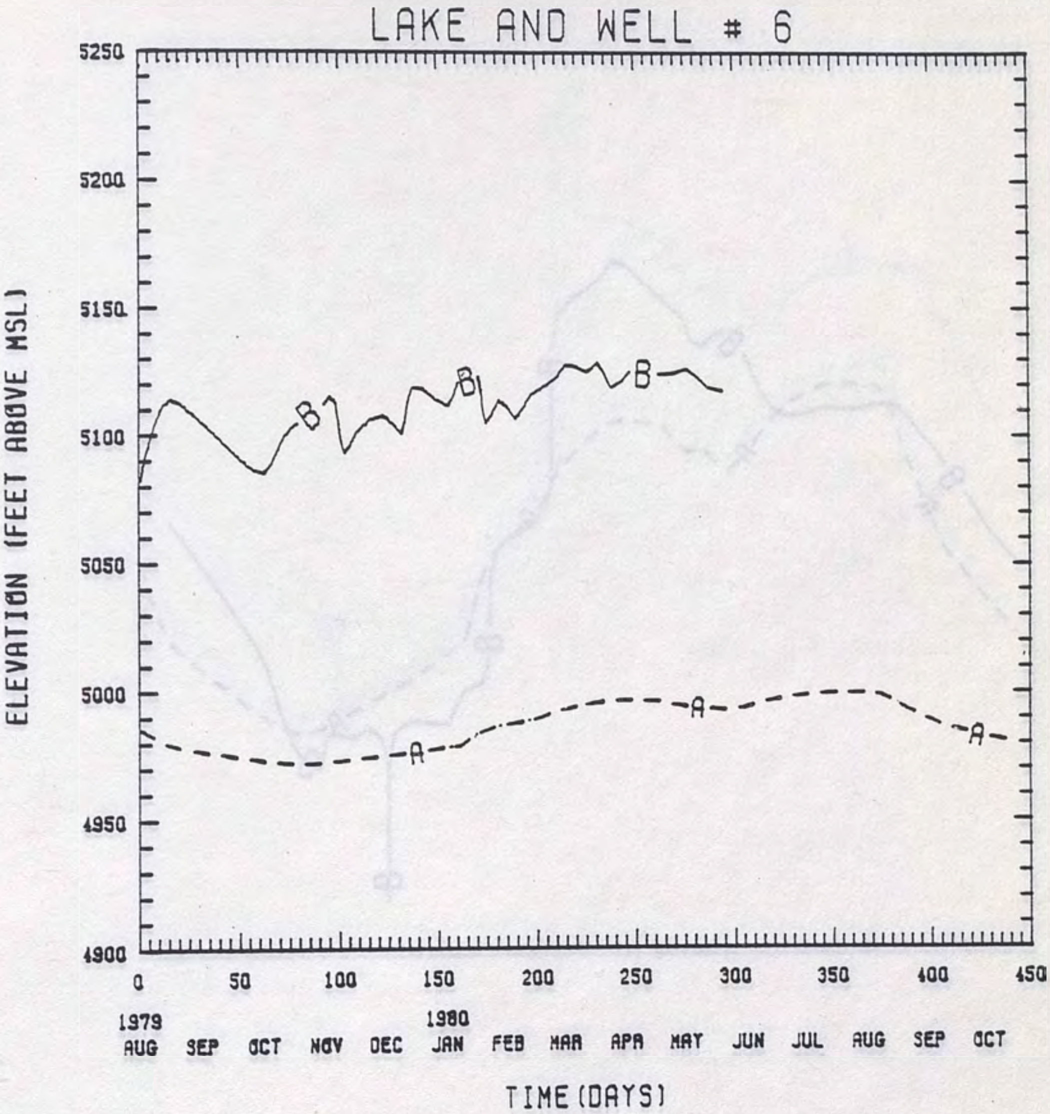


LAKE AND WELL # 3
ELEVATION (FEET ABOVE MSL)
5250
5200
5150
5100
5050
5000
4950
4900
B
A
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)

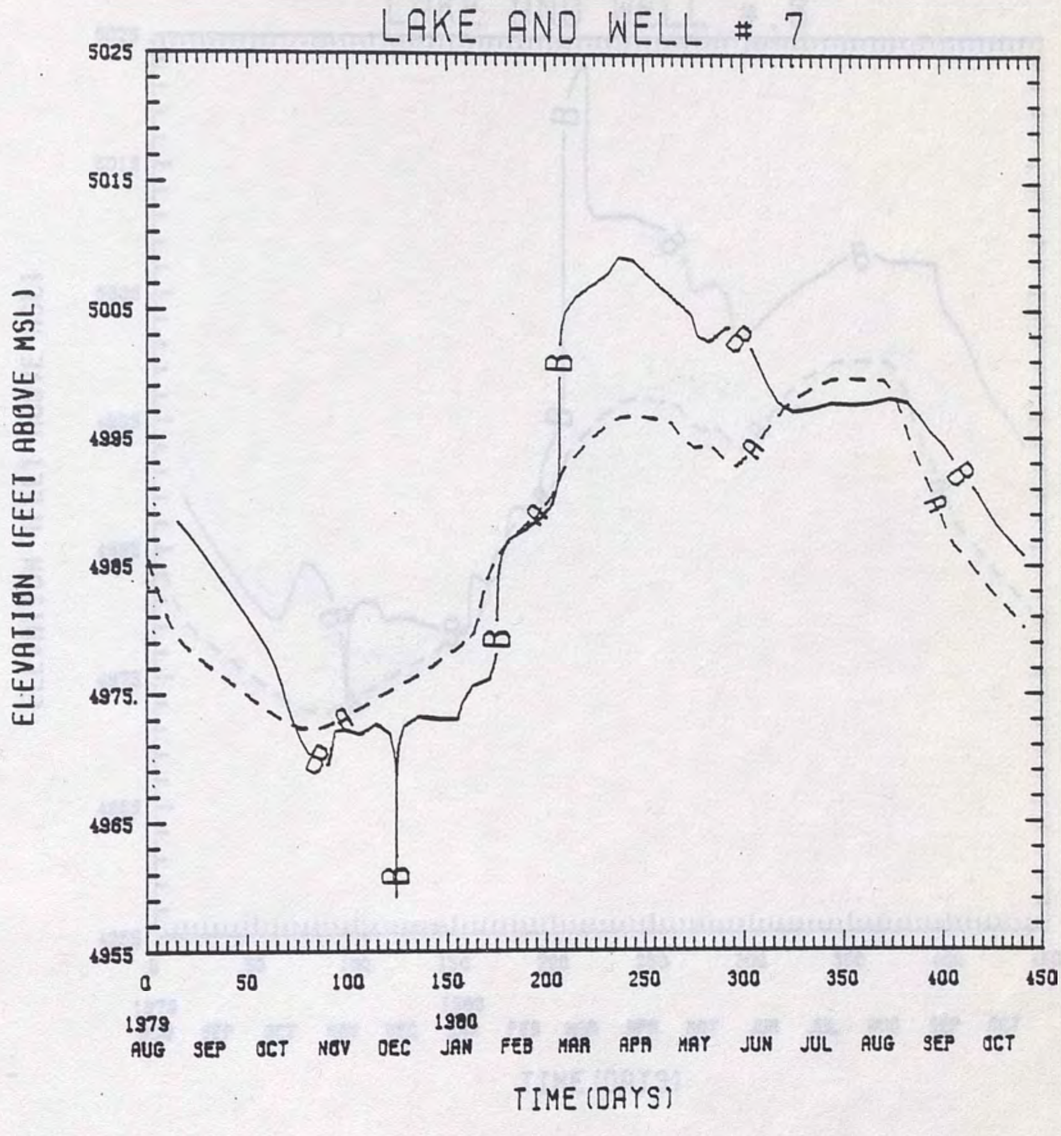


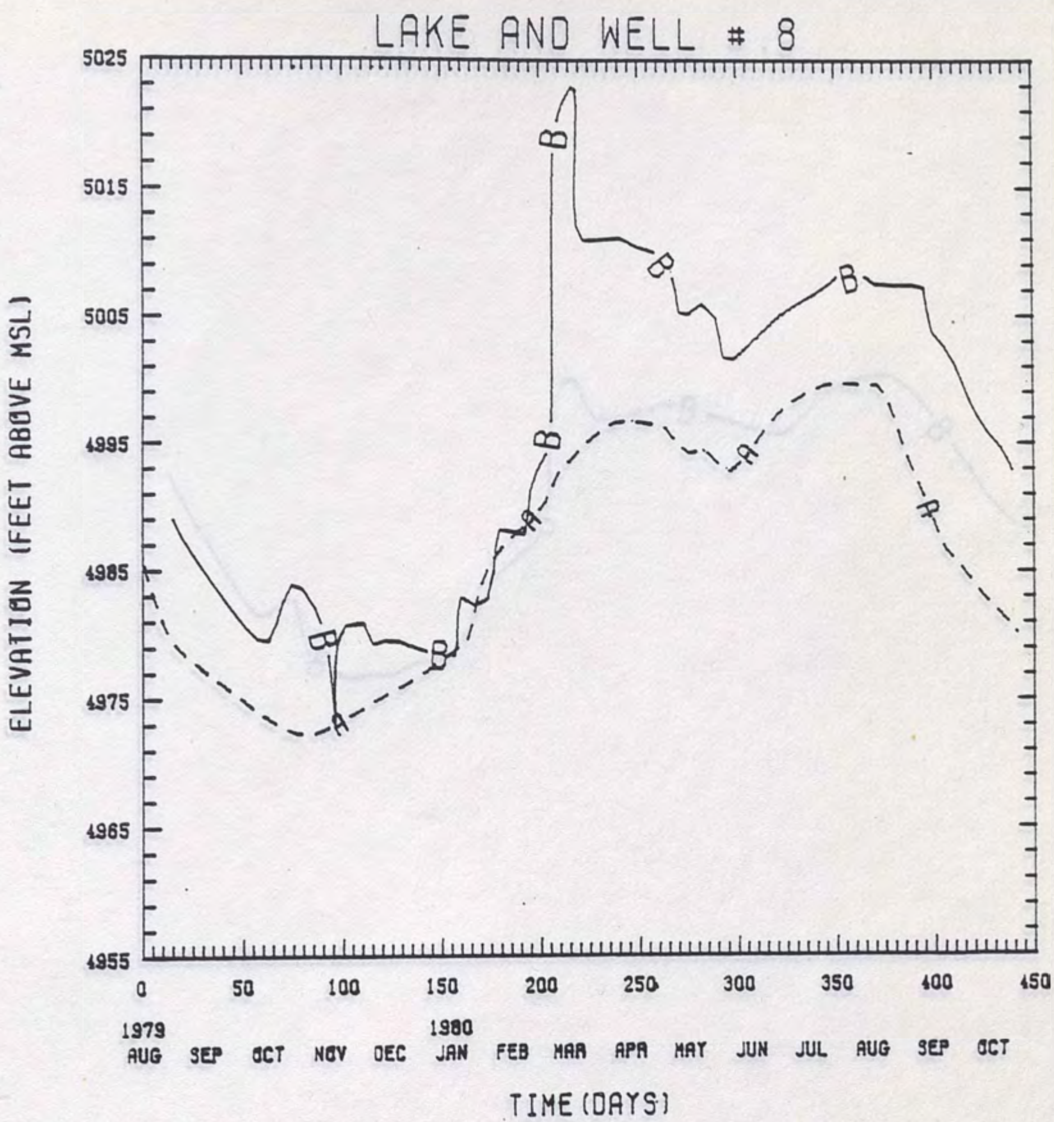
LAKE AND WELL # 4
5250
5200
5150
5100
5050
5000
4950
4900
ELEVATION (FEET ABOVE MSL)
B
A
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG SEP OCT NOV DEC JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)


LAKE AND WELL # 5
ELEVATION (FEET ABOVE MSL)
5250
5200
5150
5100
5050
5000
4950
4900
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

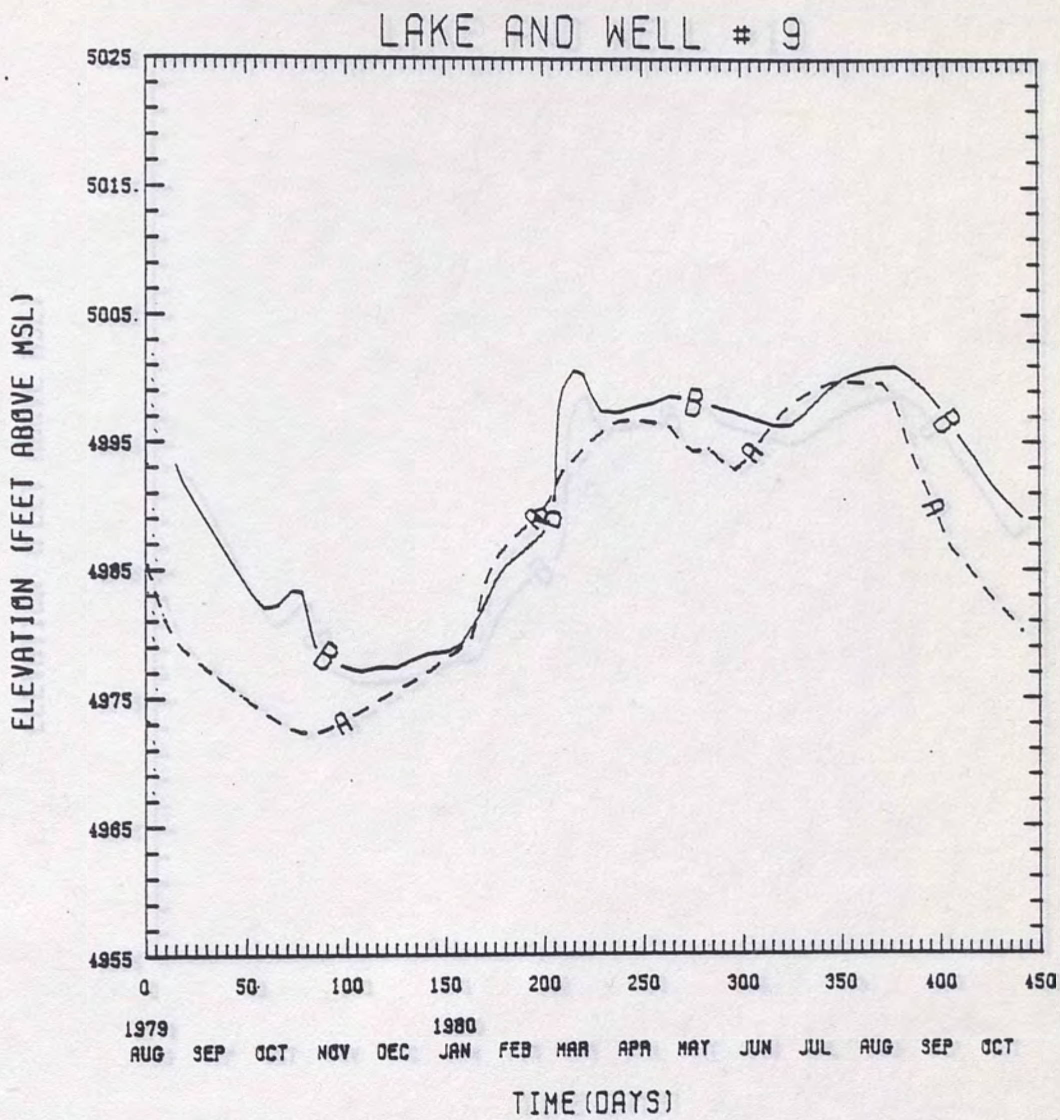


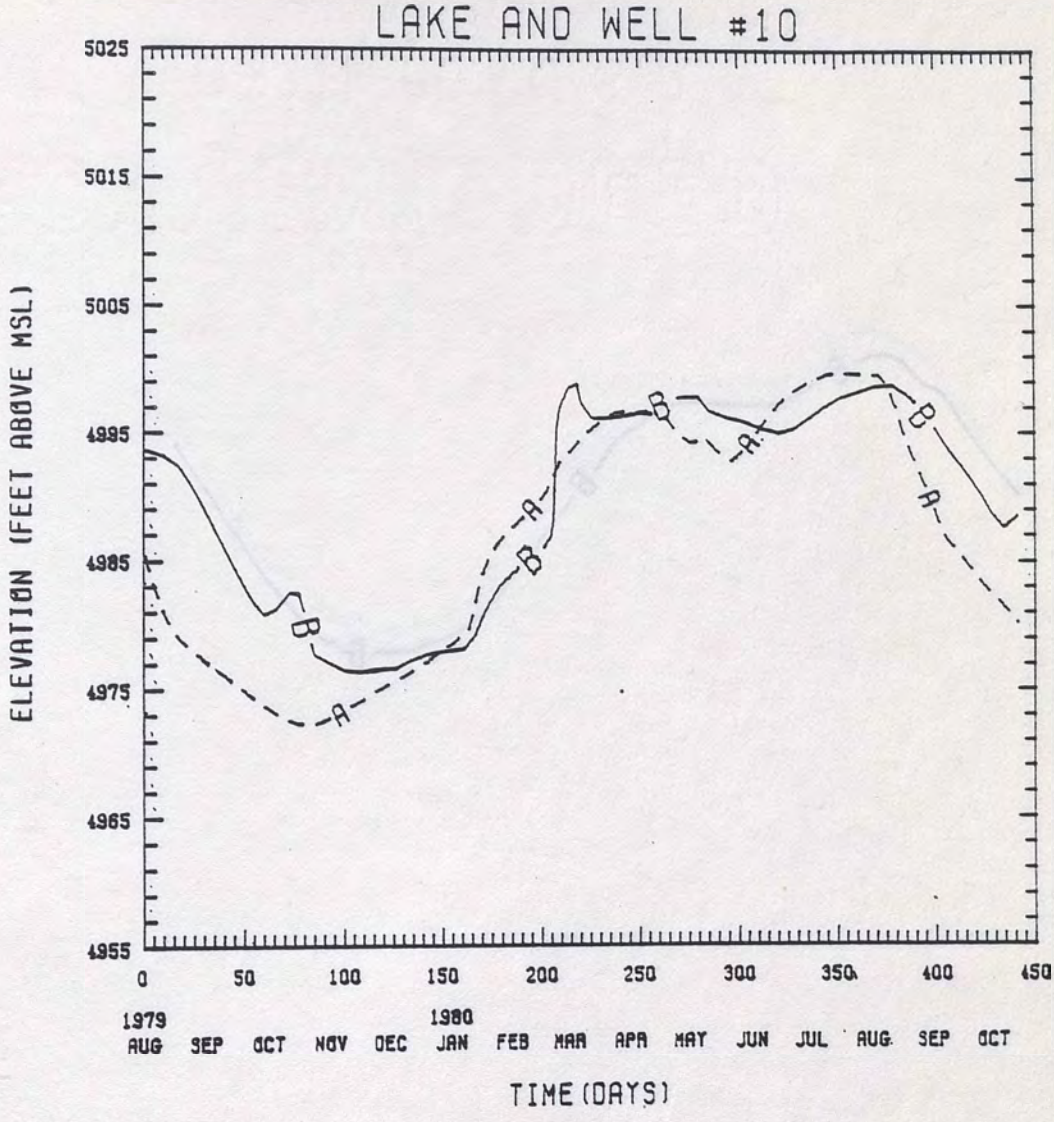
LAKE AND WELL # 6
5250
5200
5150
5100
5050
5000
4950
4900
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

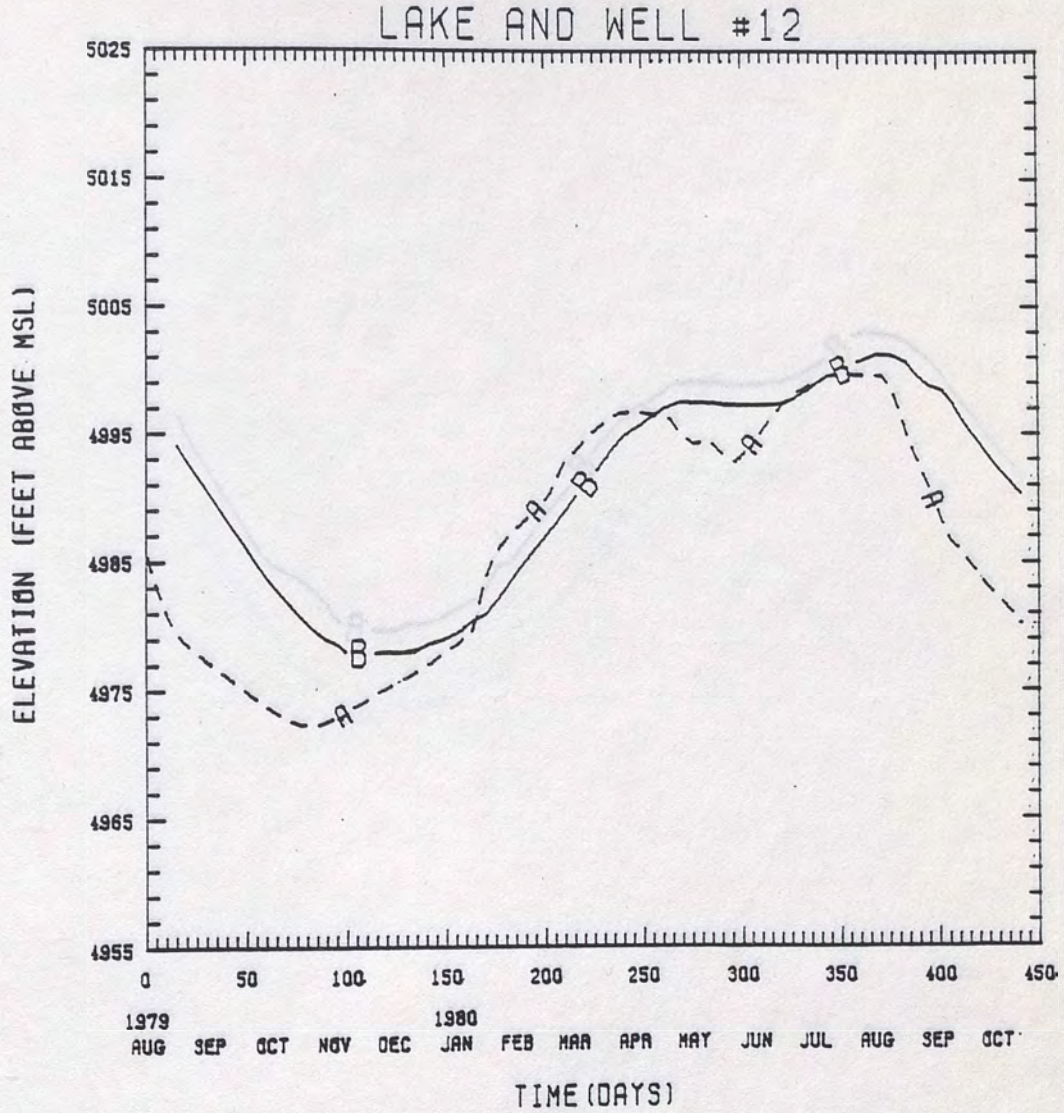

LAKE AND WELL # 7
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975.
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG
SEP
OCT
NOV
DEC
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

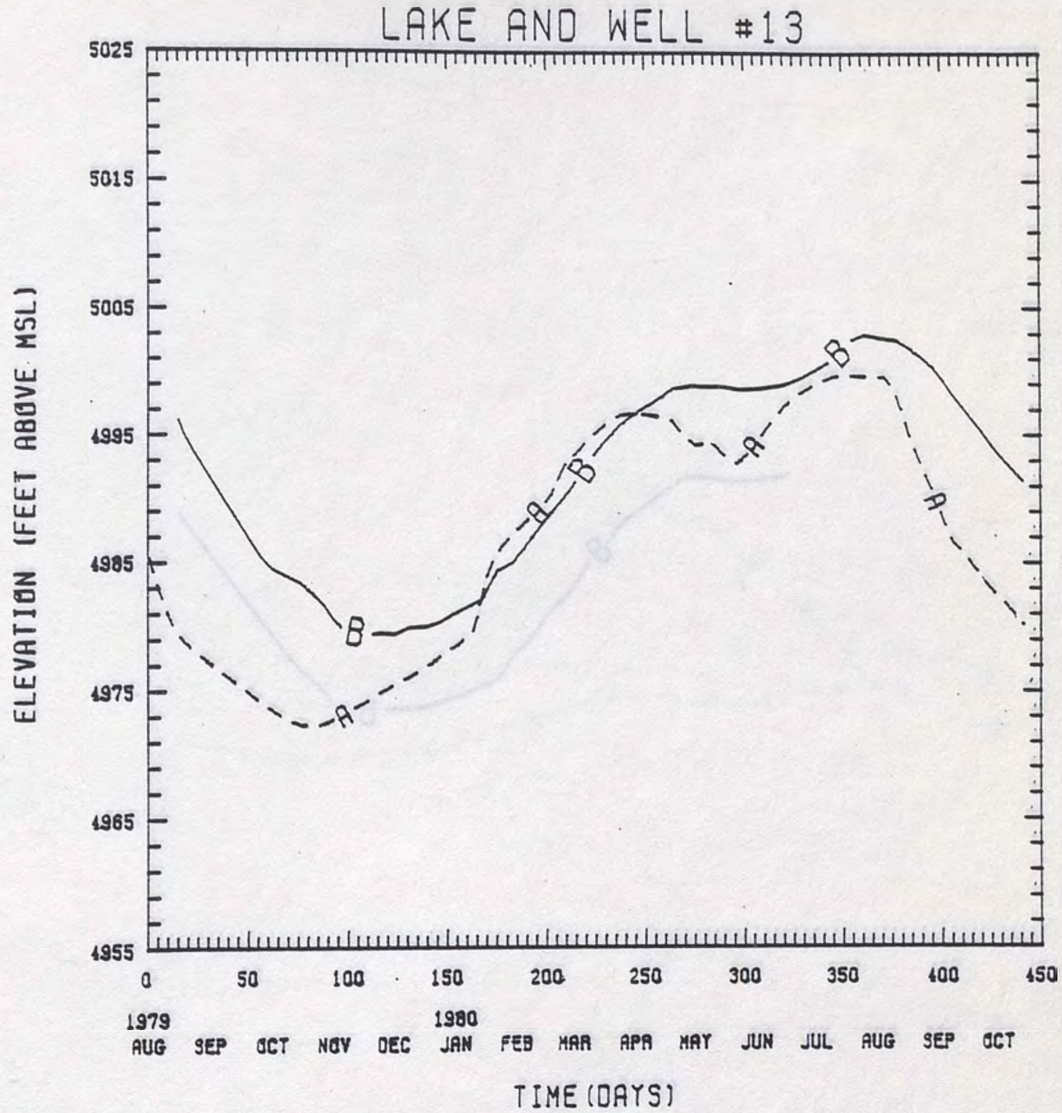


LAKE AND WELL # 8
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

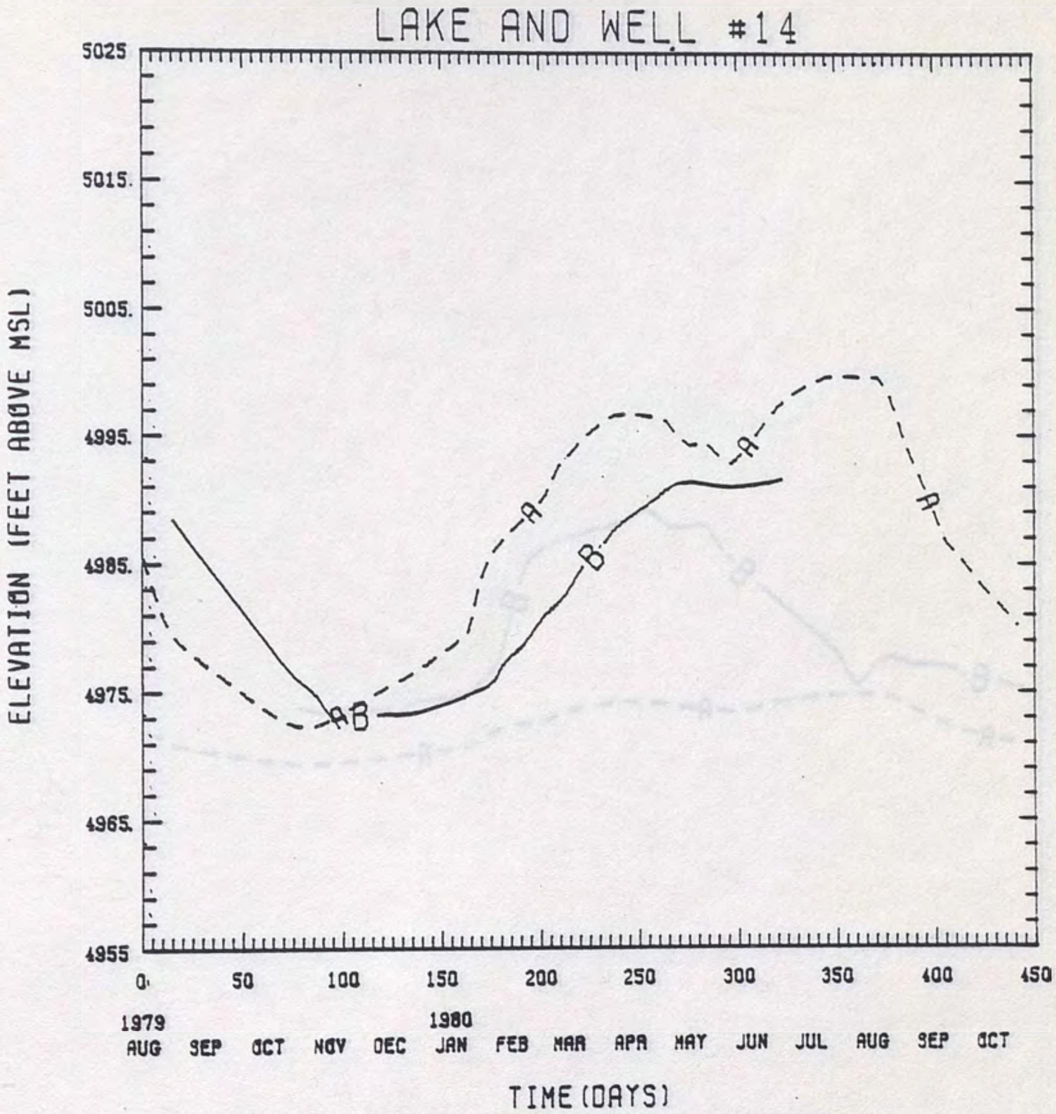

LAKE AND WELL # 9
ELEVATION (FEET ABOVE MSL)
5025
5015.
5005.
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

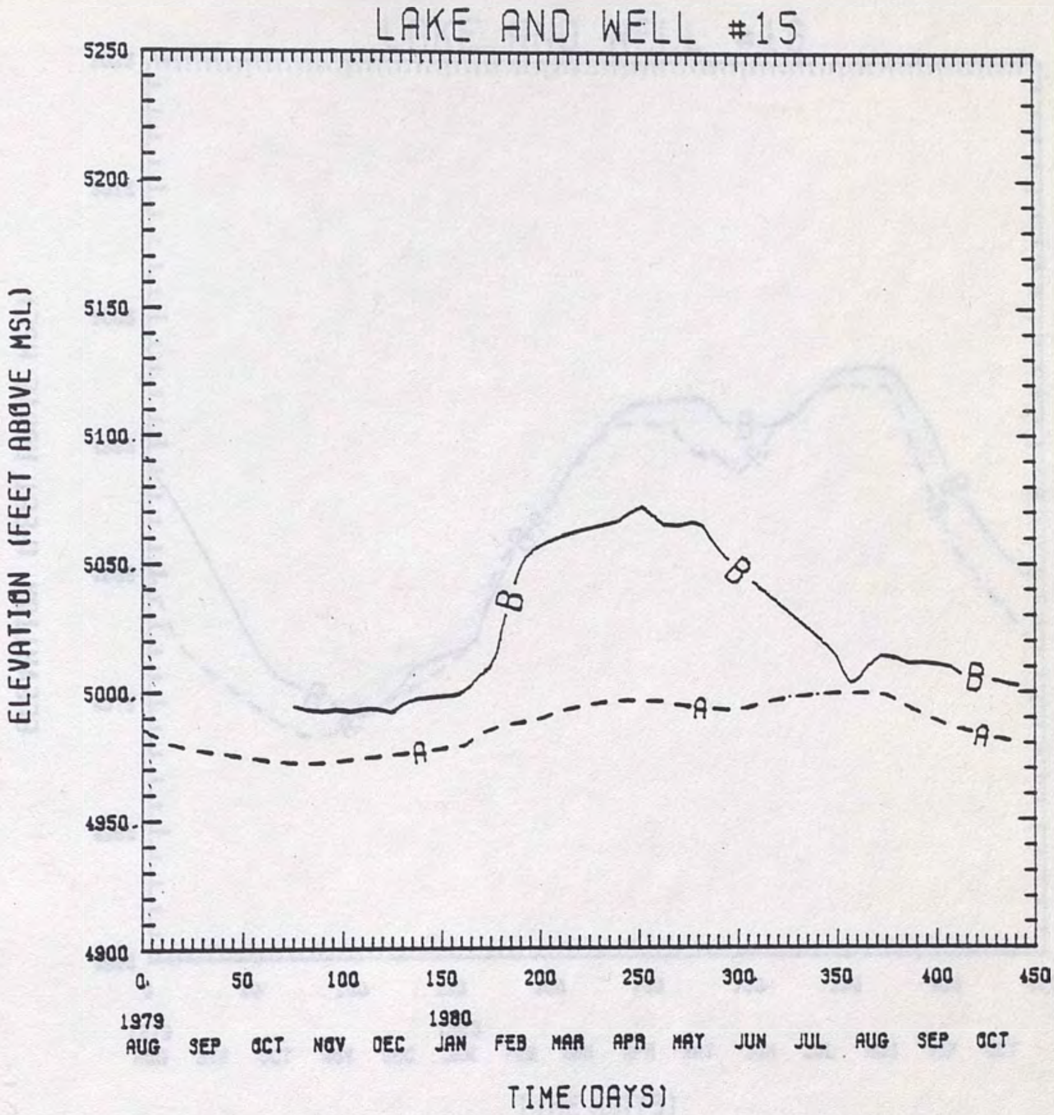

LAKE AND WELL #10
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

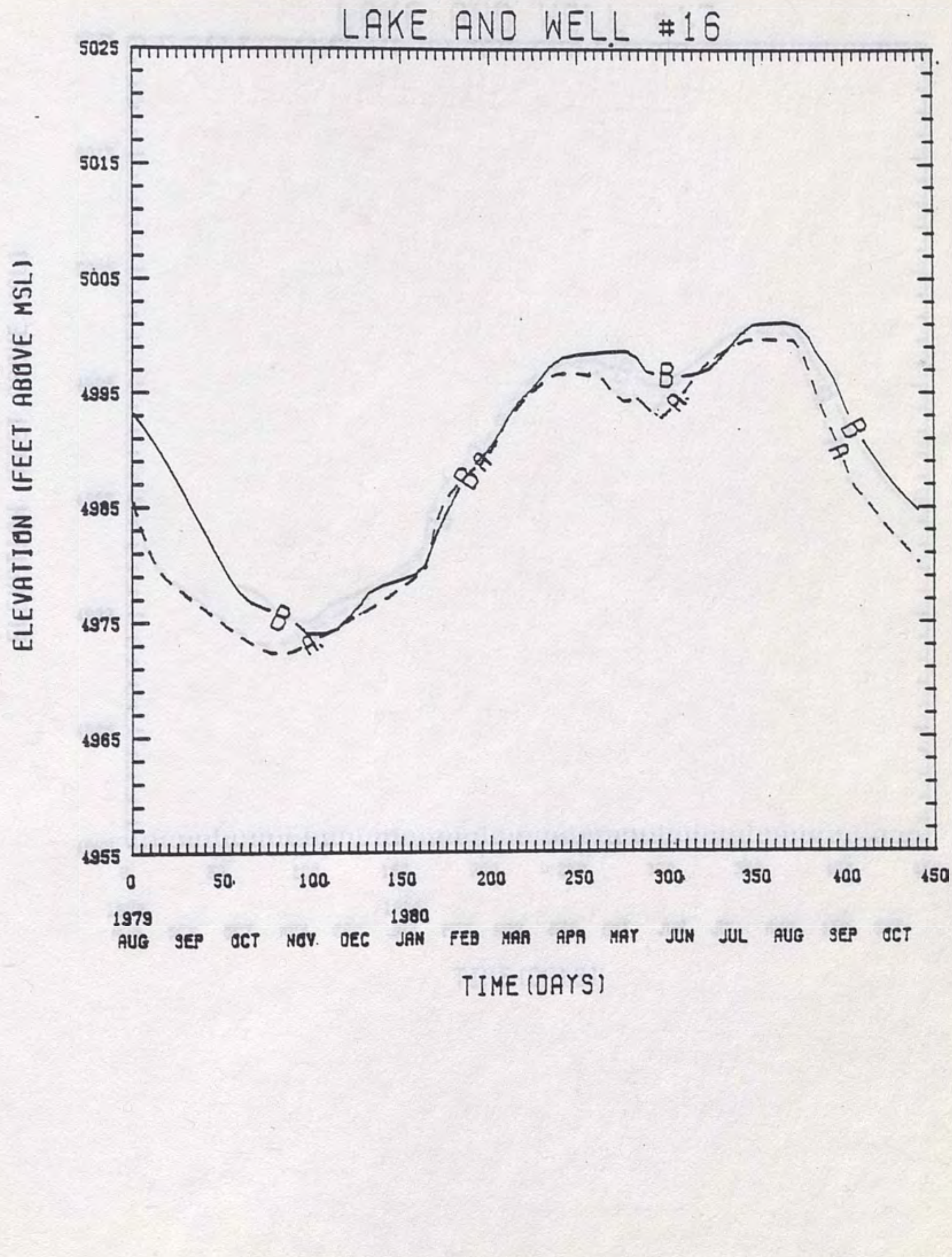

LAKE AND WELL #12
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG SEP OCT NOV DEC JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)
A
B

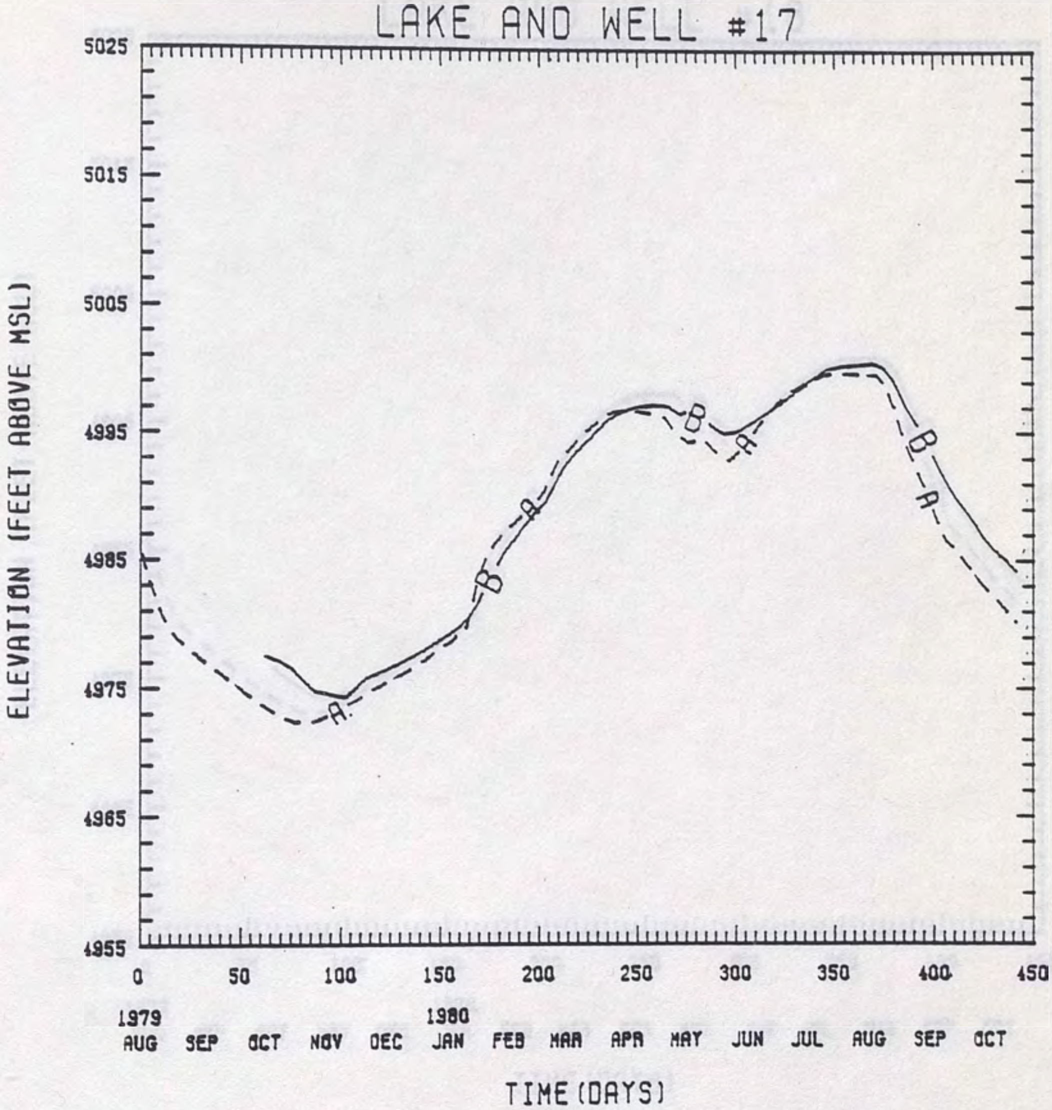


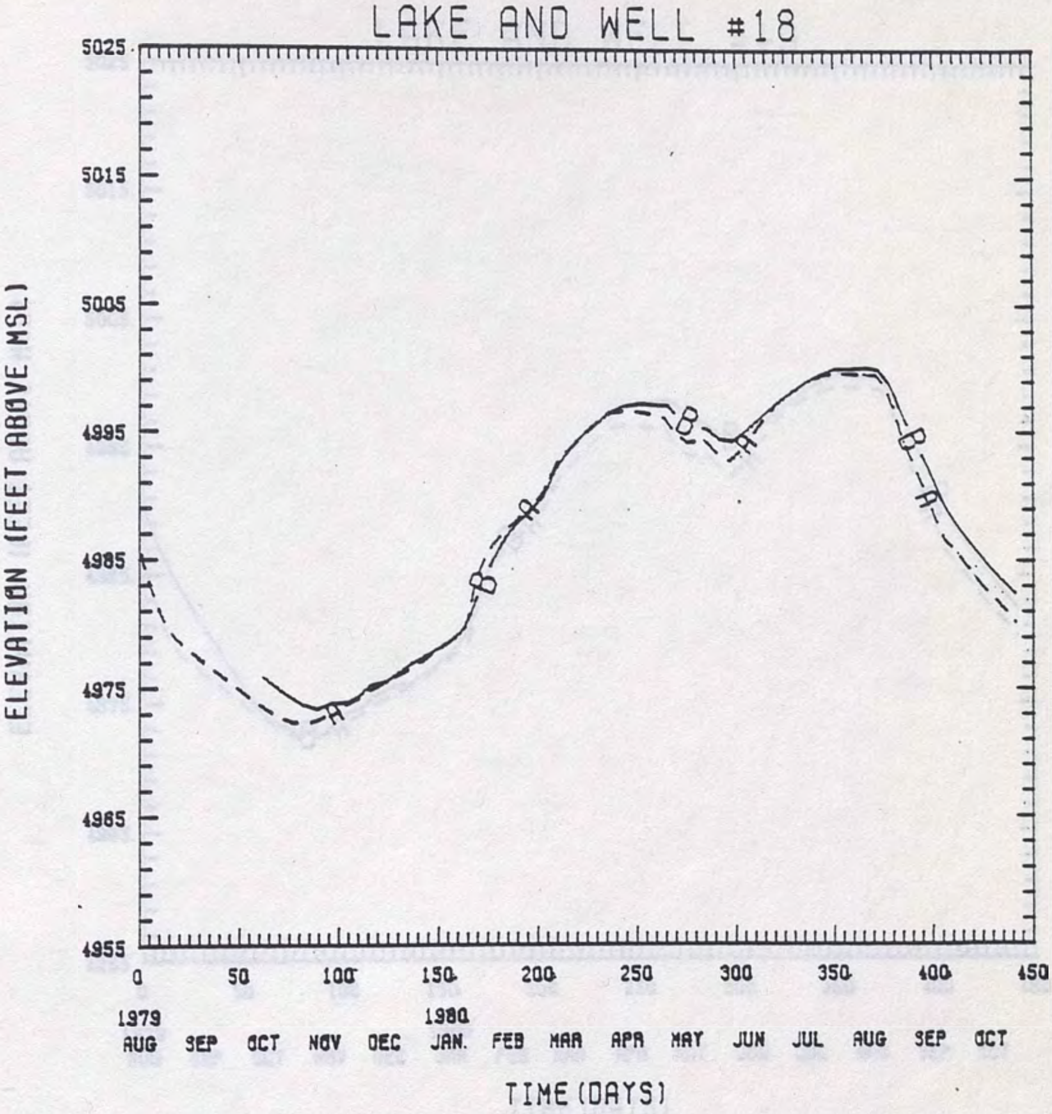
LAKE AND WELL #13
5025
5015
5005
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

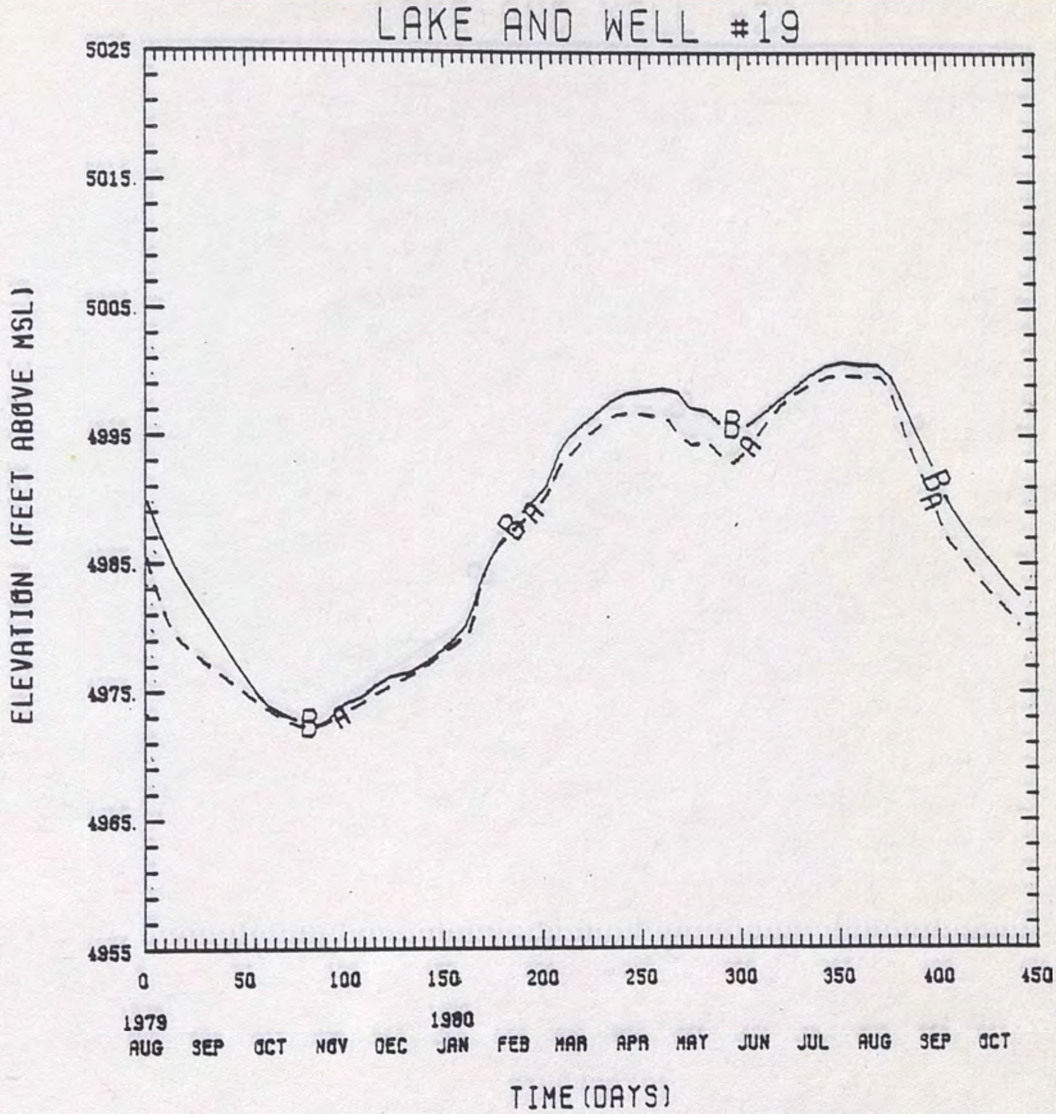

LAKE AND WELL #14
5025
5015.
5005.
4995.
4985.
4975.
4965.
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG
SEP
OCT
NOV
DEC
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

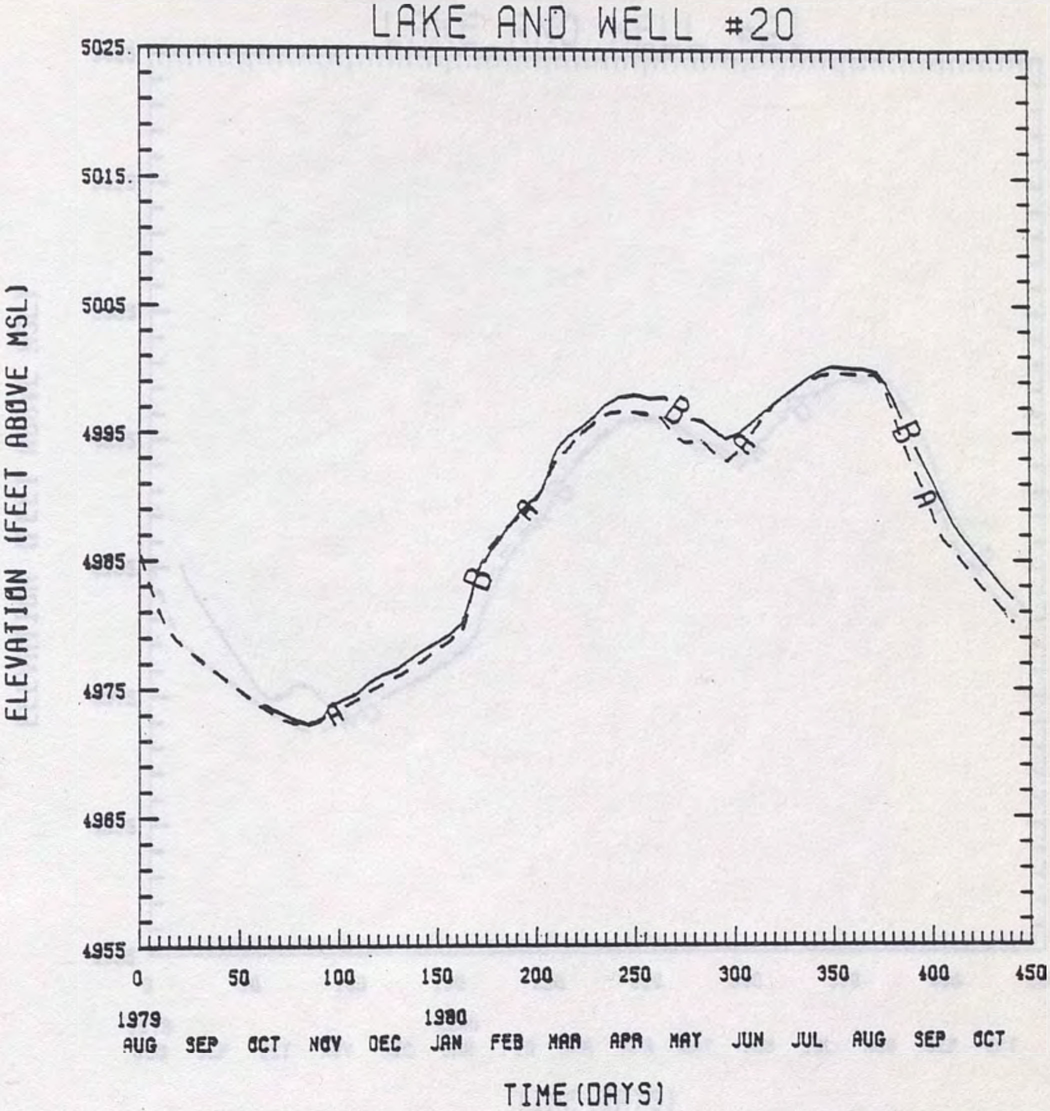

LAKE AND WELL #15
ELEVATION (FEET ABOVE MSL)
5250
5200
5150
5100.
5050.
5000.
4950.
4900
0.
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

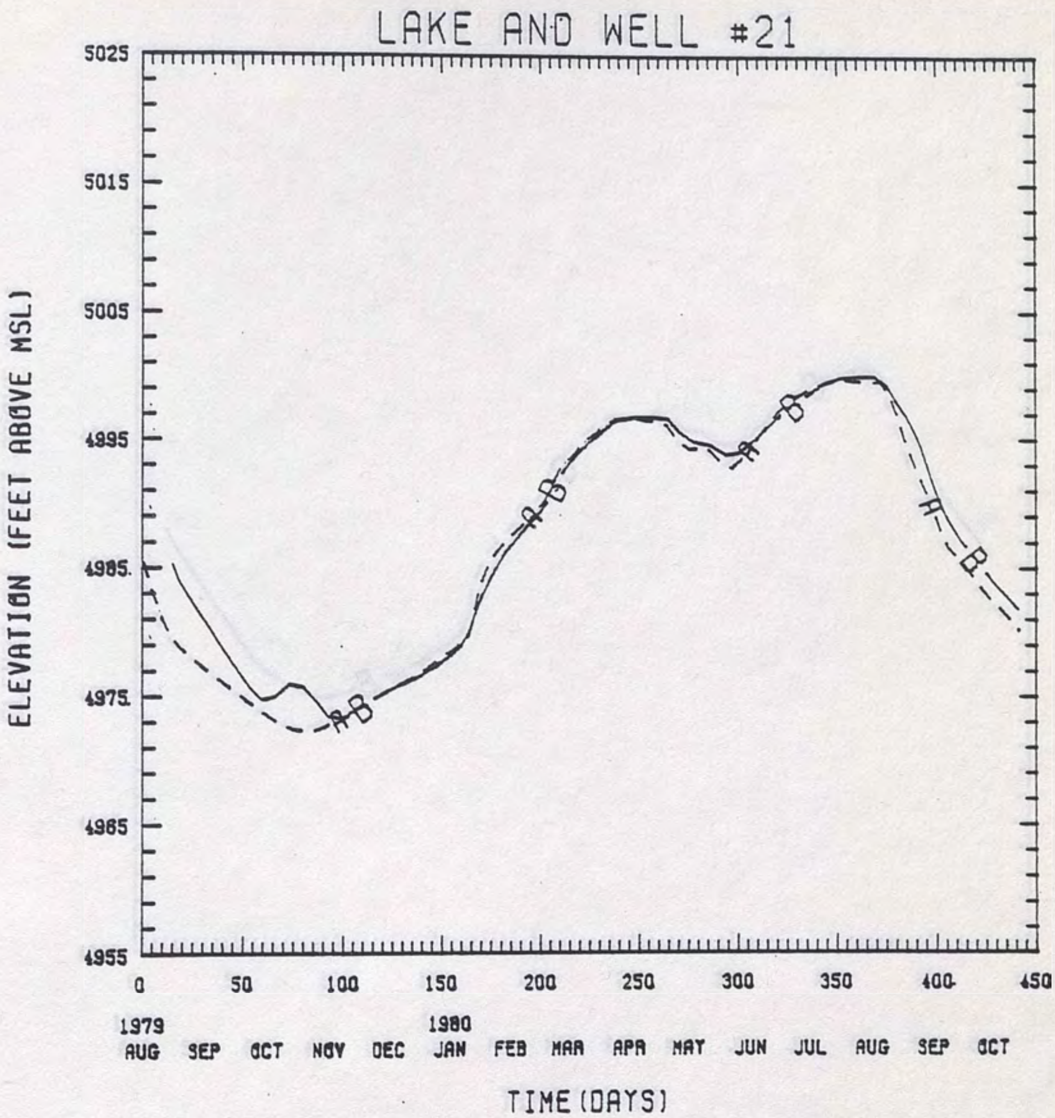

LAKE AND WELL #16
5025
5015
5005
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
A
B
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)

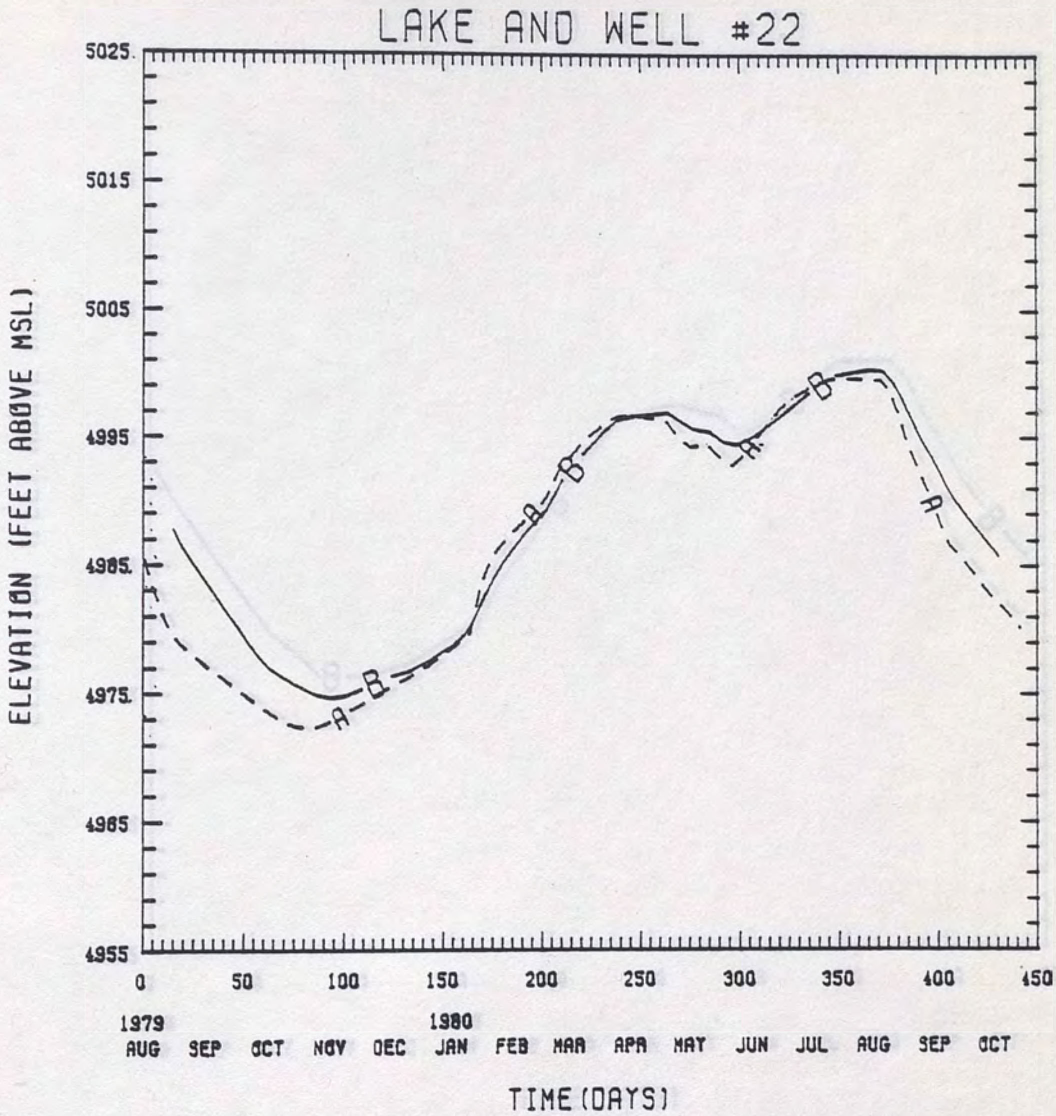

LAKE AND WELL #17
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B


LAKE AND WELL #18
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG
SEP
OCT
NOV
DEC
JAN.
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B


LAKE AND WELL #19
ELEVATION (FEET ABOVE MSL)
5025
5015.
5005.
4995
4985.
4975.
4965.
4955
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

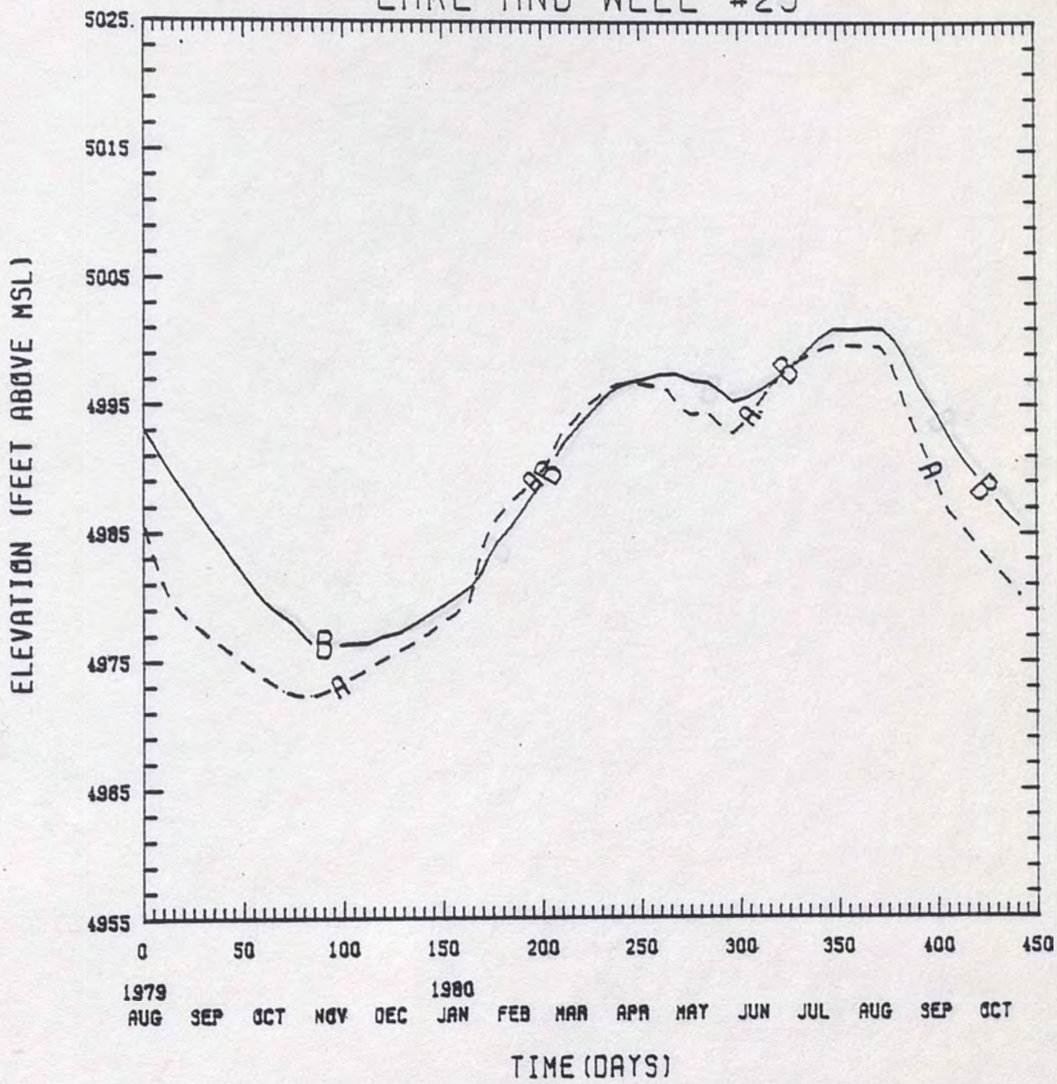

LAKE AND WELL #20
5025.
5015.
5005
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

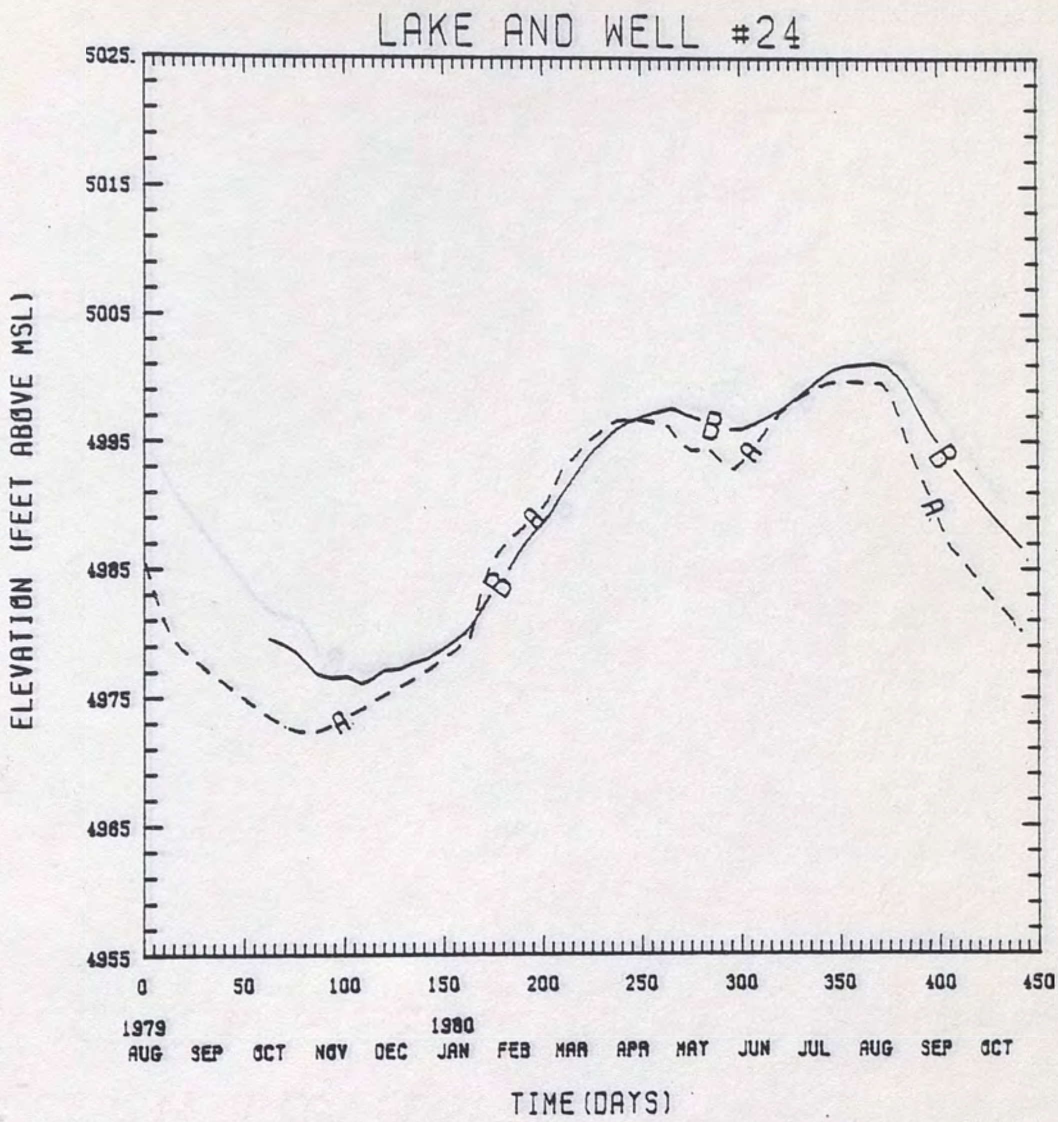

LAKE AND WELL #21
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985.
4975.
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

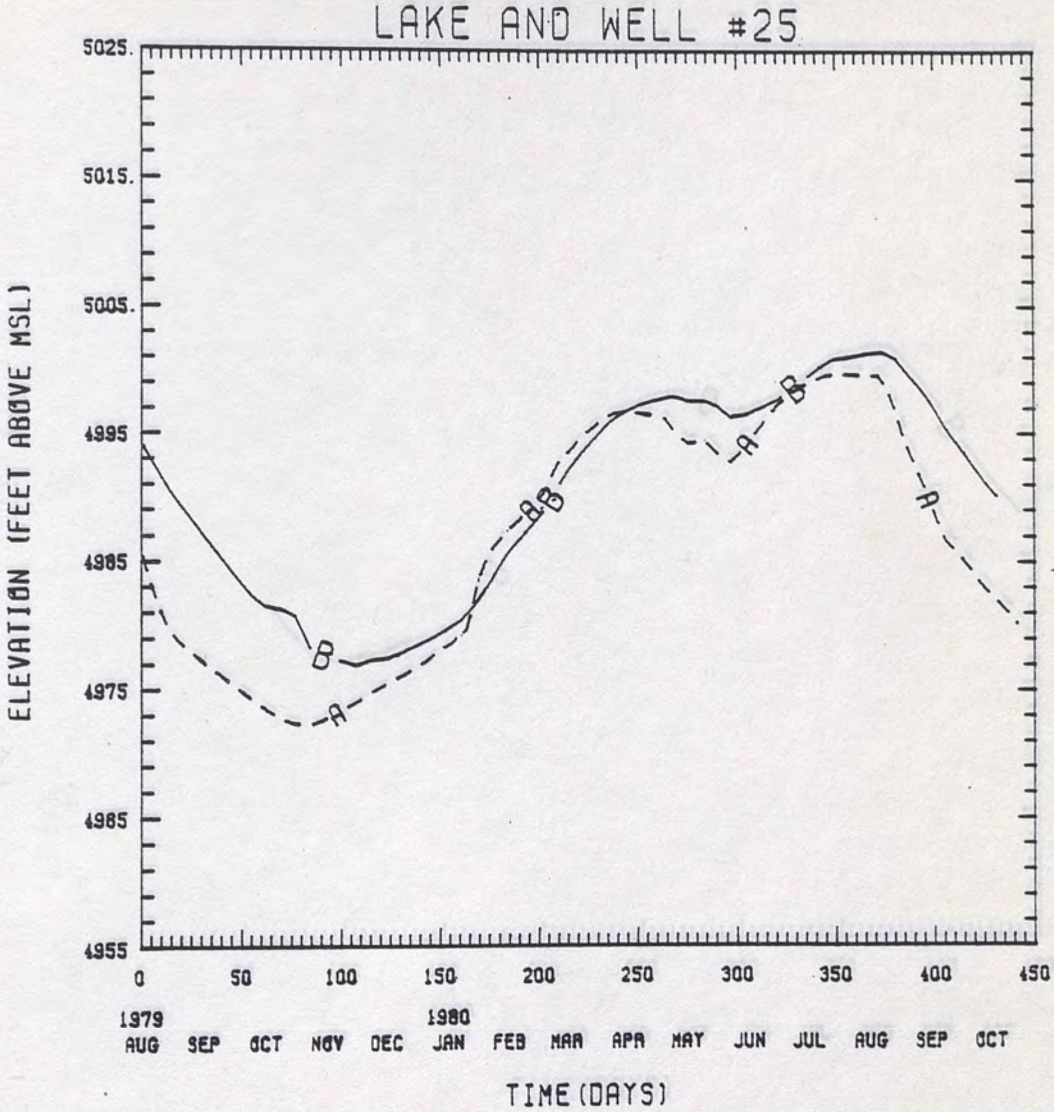


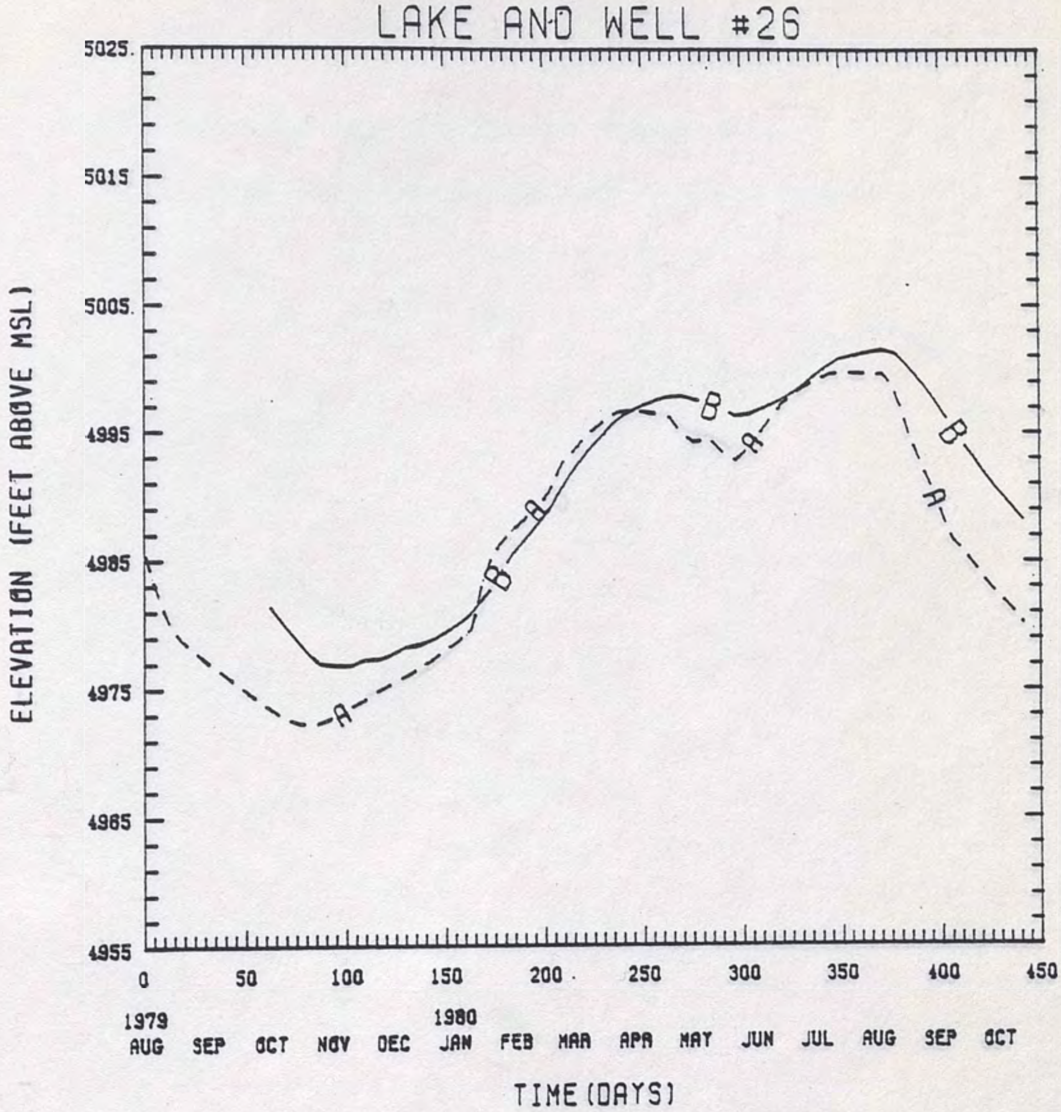
LAKE AND WELL #22
5025
5015
5005
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG
SEP
OCT
NOV
DEC
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

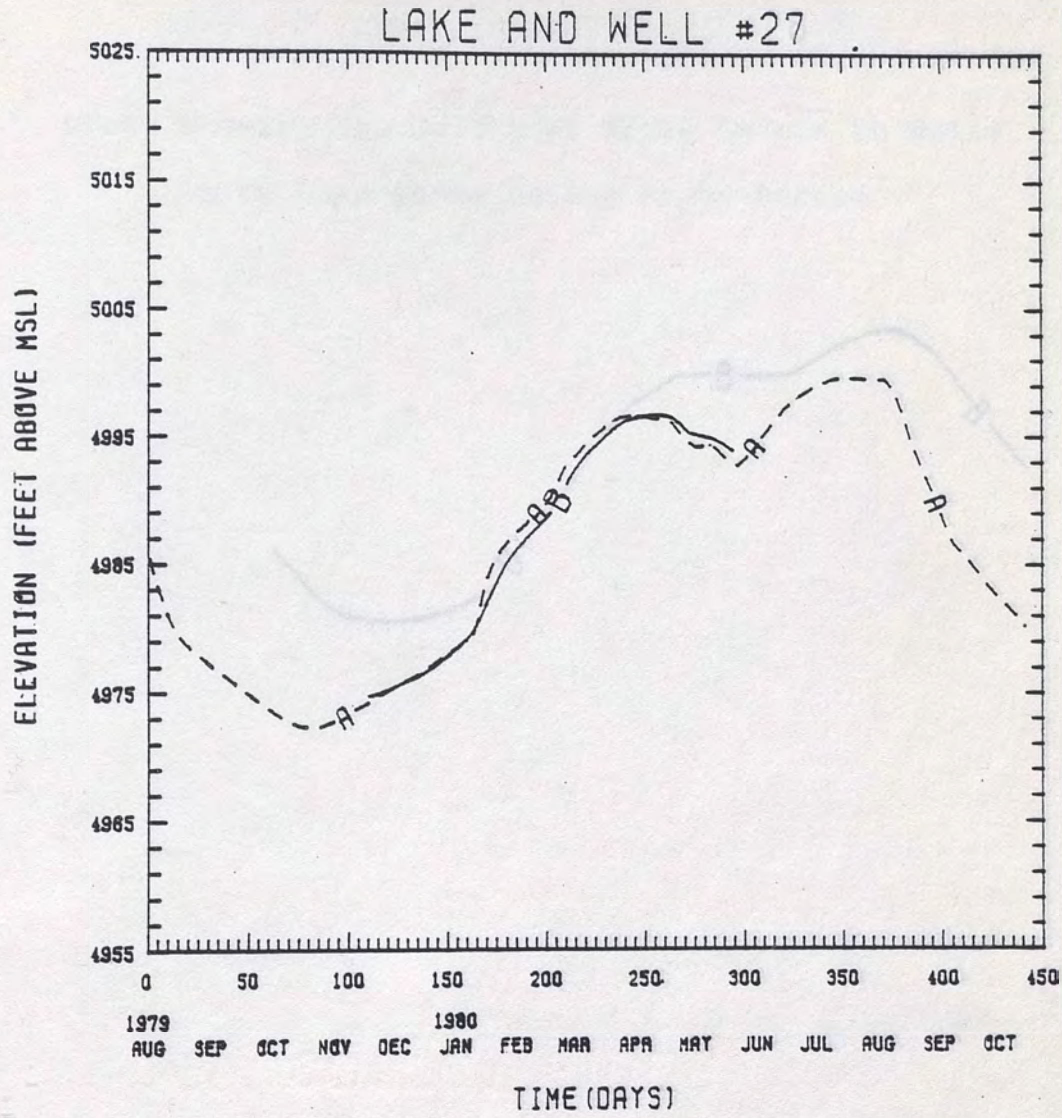
LAKE AND WELL #23

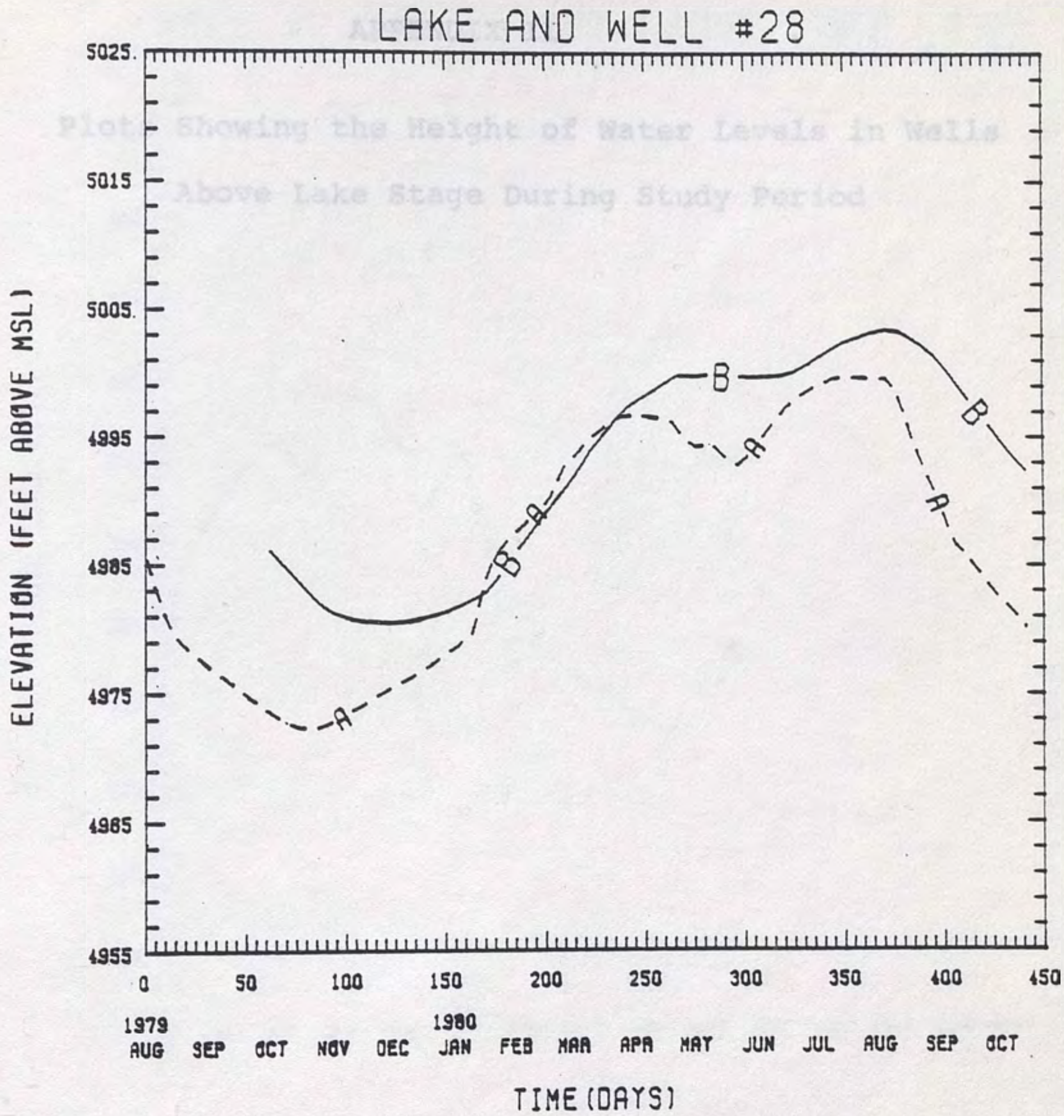



LAKE AND WELL #24
5025.
5015
5005
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG SEP OCT NOV DEC JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)
A
B


LAKE AND WELL #25
5025.
5015.
5005.
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG SEP OCT NOV DEC JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)
A
B

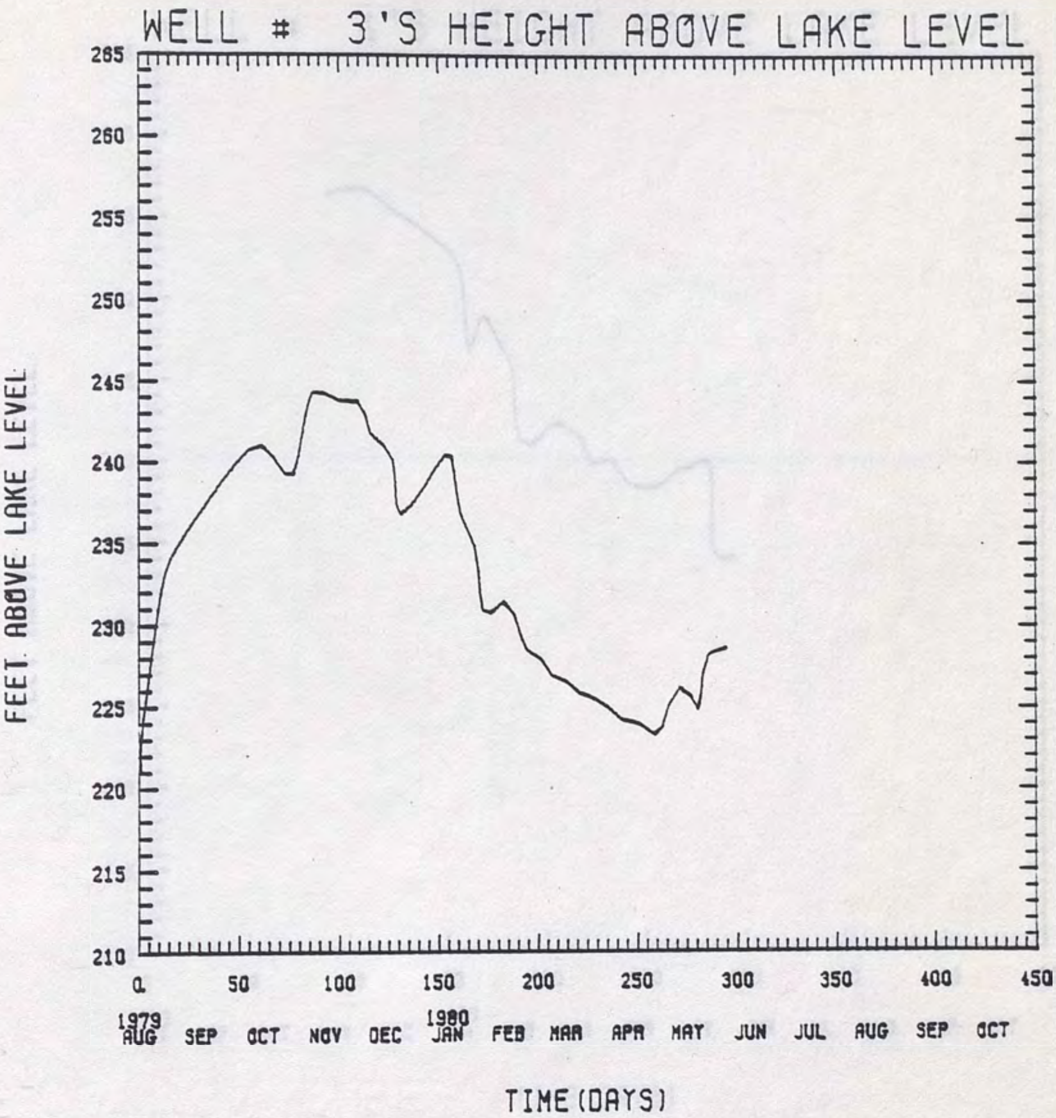

LAKE AND WELL #26
ELEVATION (FEET ABOVE MSL)
5025.
5015
5005.
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG
SEP
OCT
NOV
DEC
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

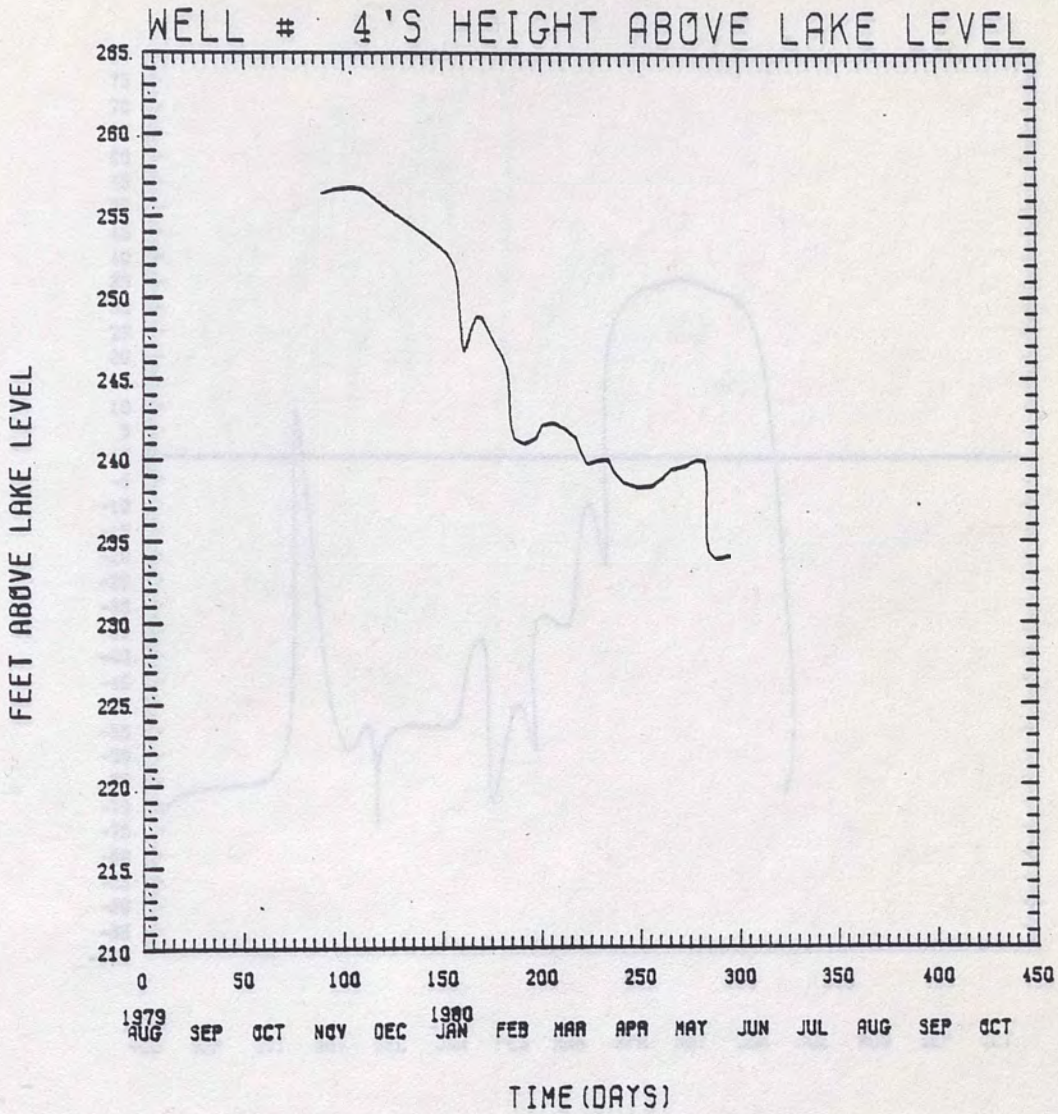

LAKE AND WELL #27
ELEVATION (FEET ABOVE MSL)
5025
5015
5005
4995
4985
4975
4965
4955
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG SEP OCT NOV DEC JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)
A
B

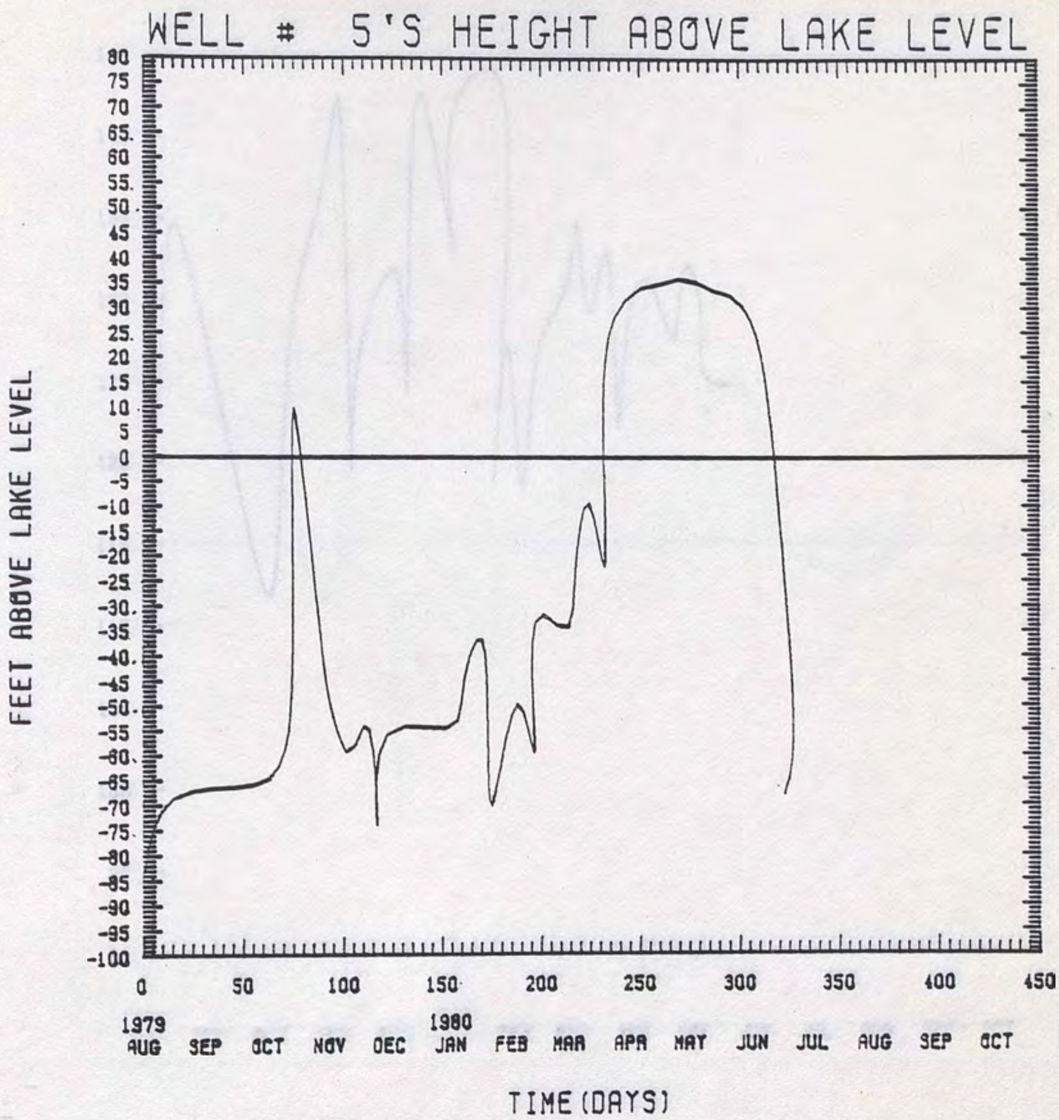

LAKE AND WELL #28
5025.
5015
5005.
4995
4985
4975
4965
4955
ELEVATION (FEET ABOVE MSL)
0
50
100
150
200
250
300
350
400
450
1979
1980
AUG
SEP
OCT
NOV
DEC
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)
A
B

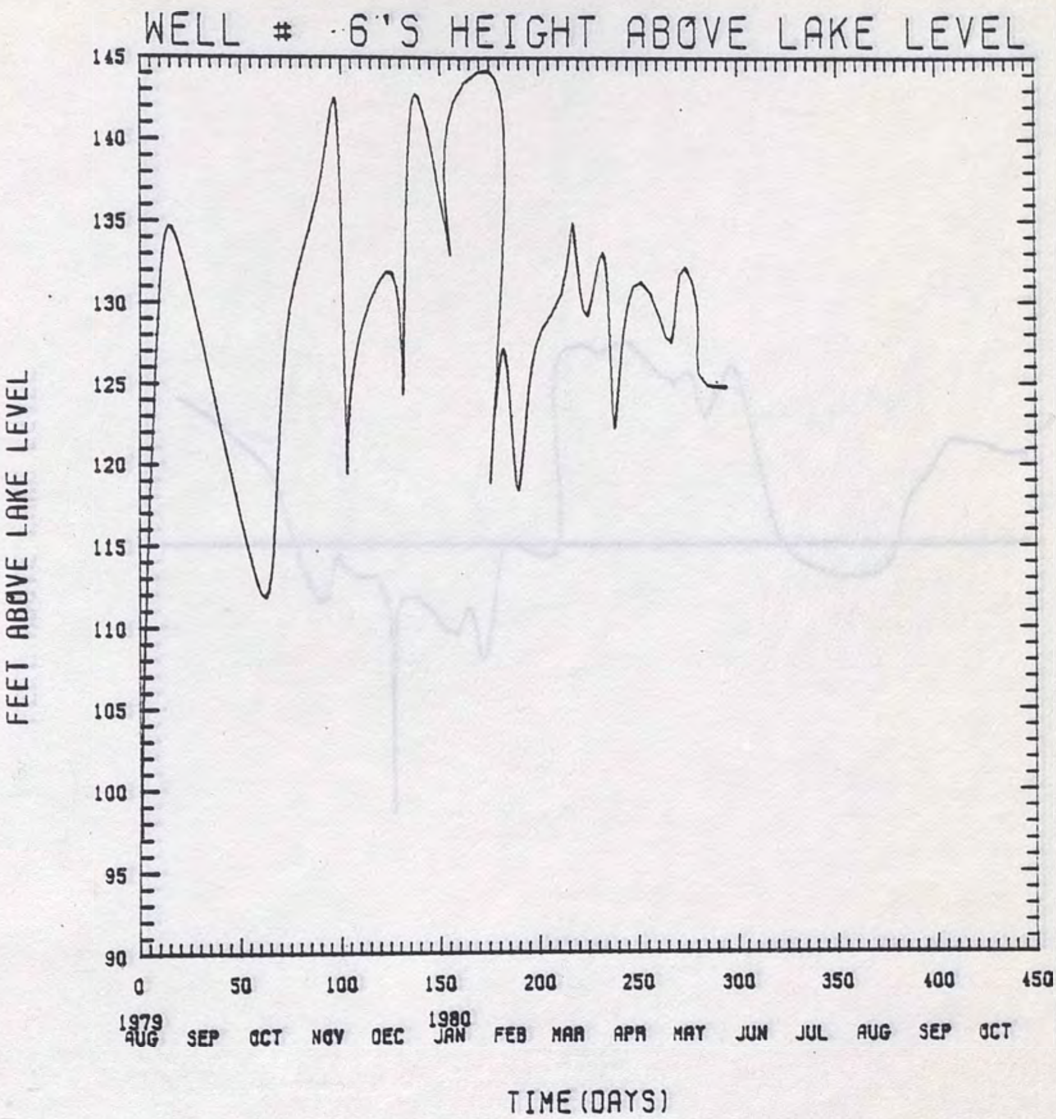
APPENDIX II

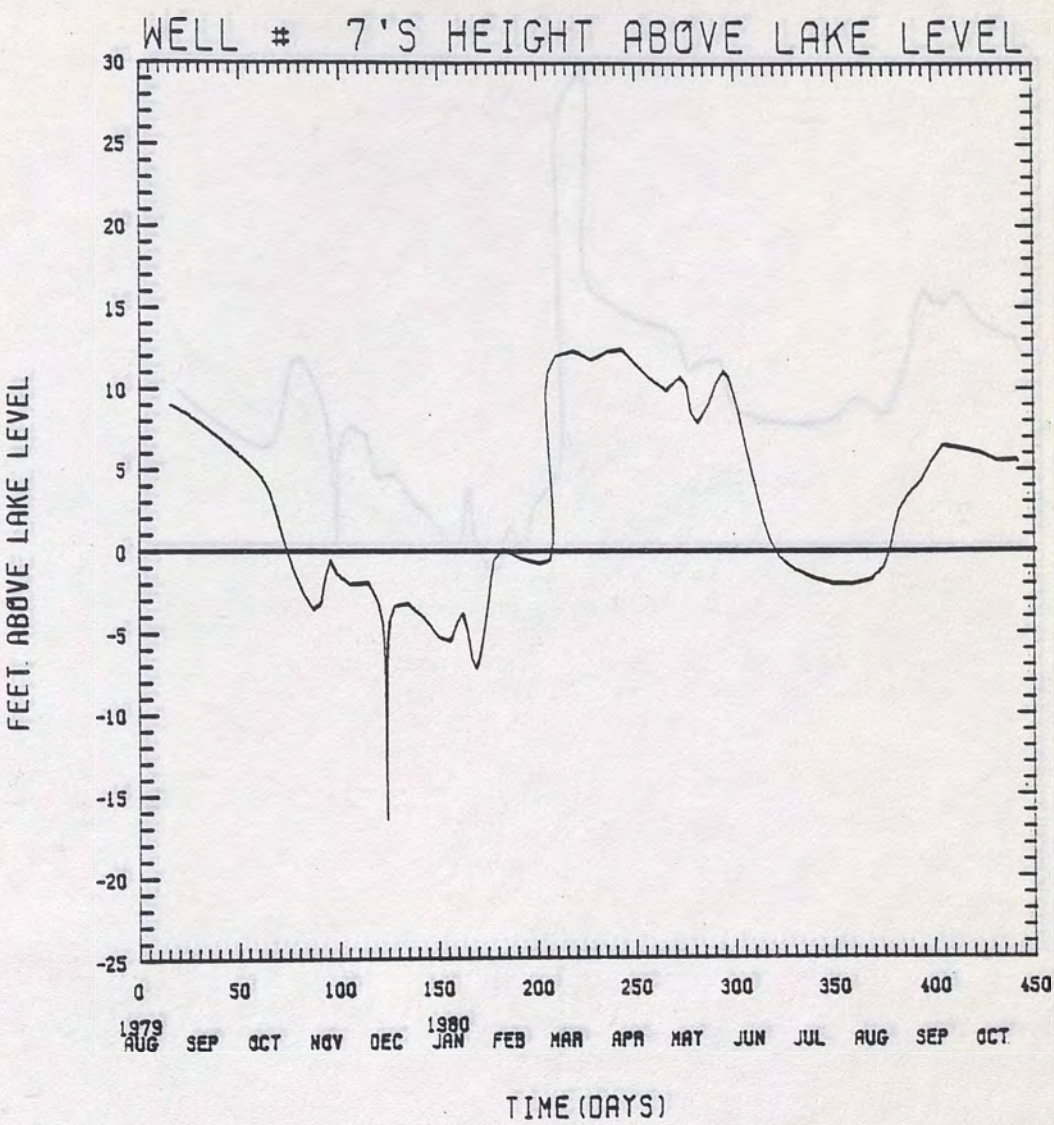
Plots Showing the Height of Water Levels in Wells Above Lake Stage During Study Period


WELL # 3'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
265
260
255
250
245
240
235
230
225
220
215
210
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)

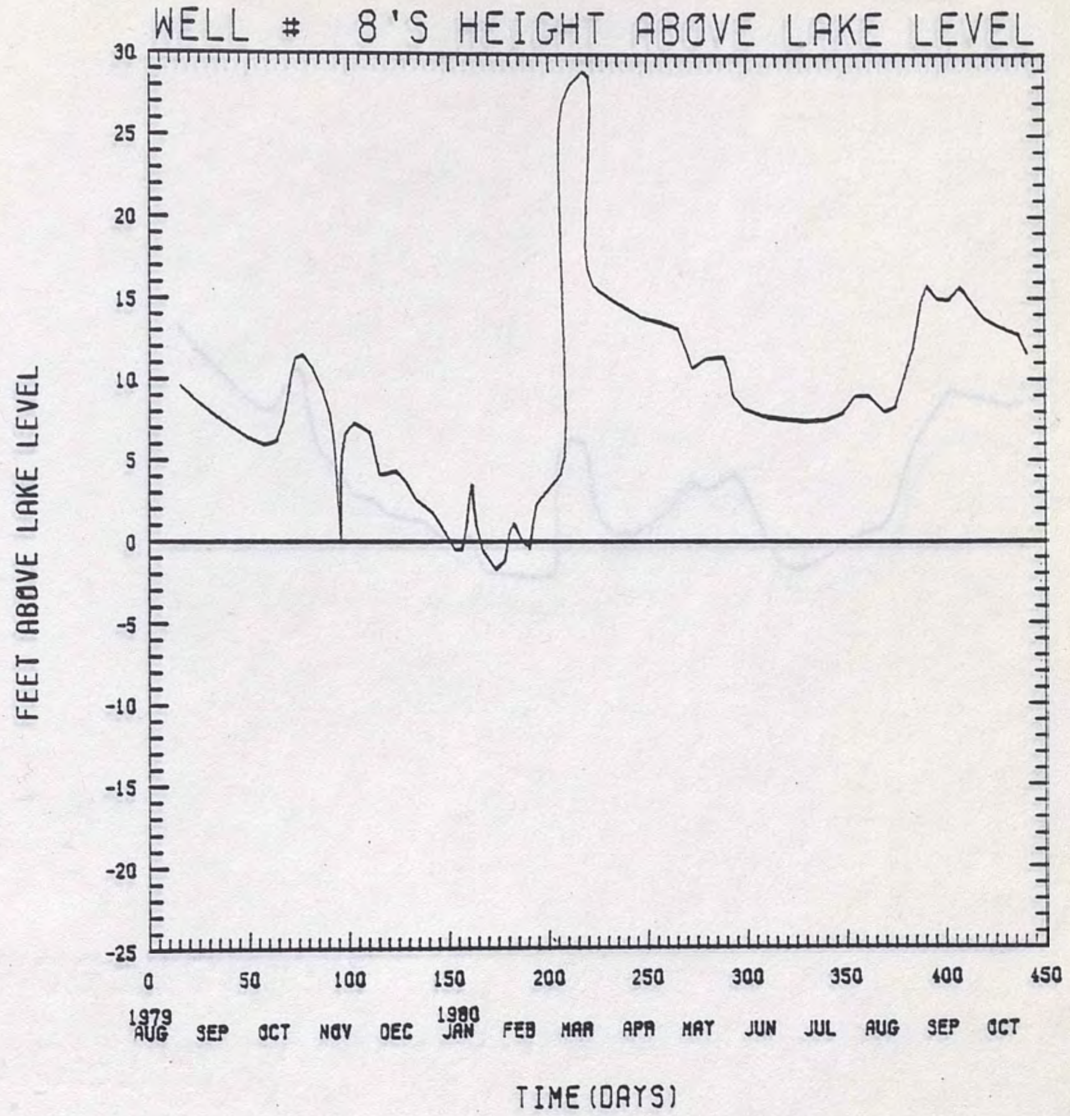

WELL # 4'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
265.
260
255
250
245.
240
235
230
225.
220
215
210
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)

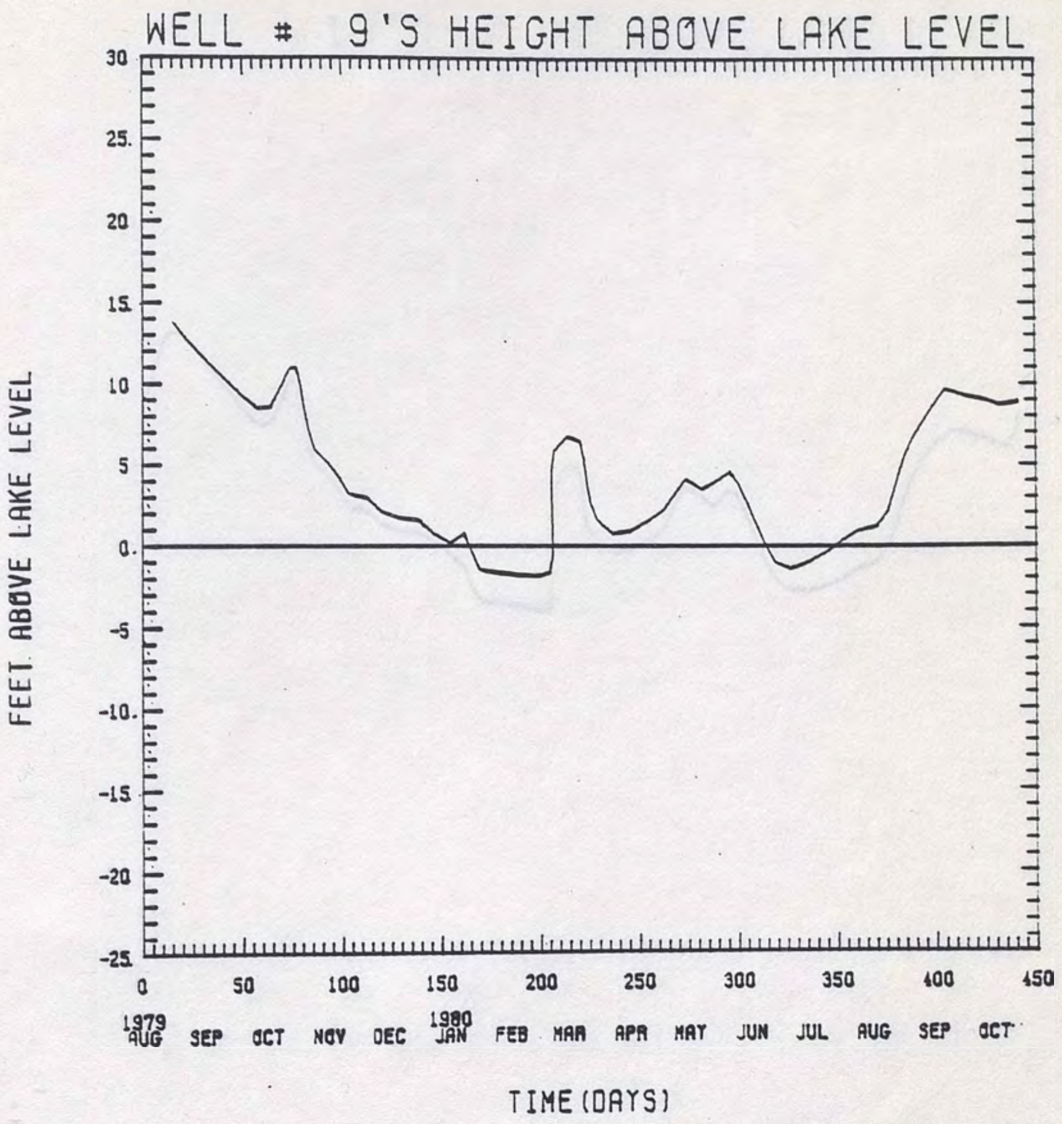

WELL # 5'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
80
75
70
65
60
55
50
45
40
35
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
-30
-35
-40
-45
-50
-55
-60
-65
-70
-75
-80
-85
-90
-95
-100
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 6'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
145
140
135
130
125
120
115
110
105
100
95
90
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)

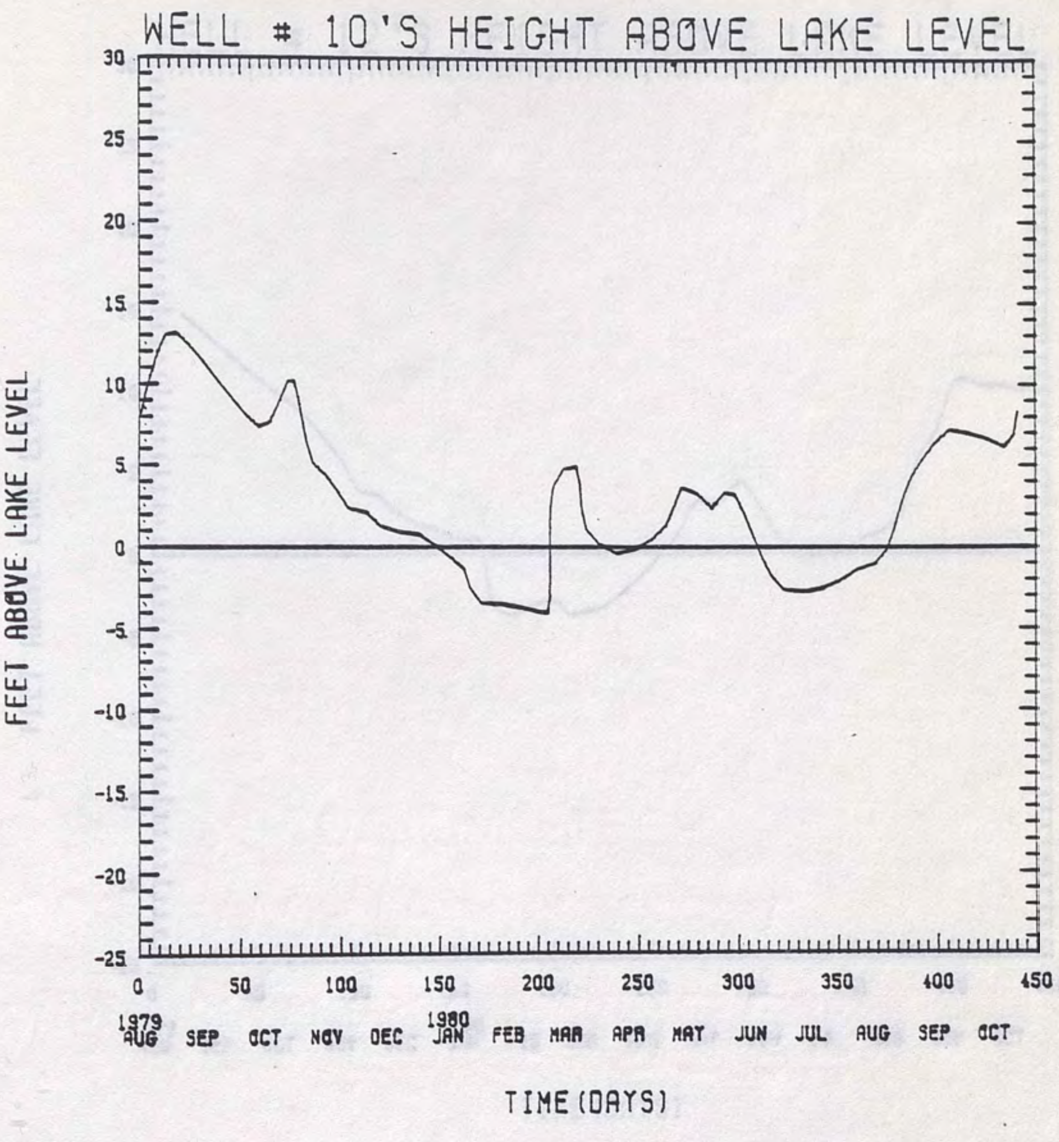


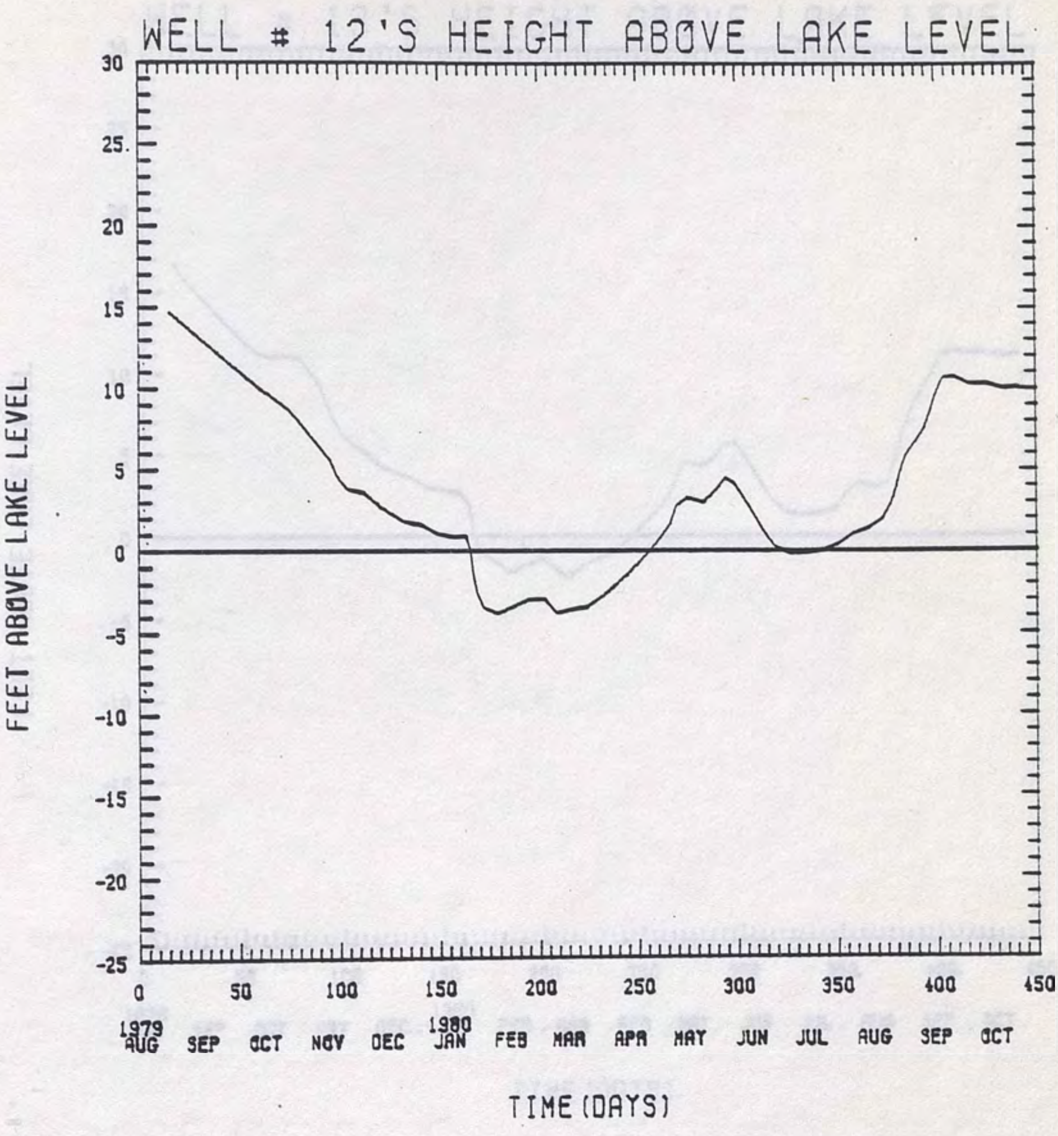
WELL # 7'S HEIGHT ABØVE LAKE LEVEL
FEET ABØVE LAKE LEVEL
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 8'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)

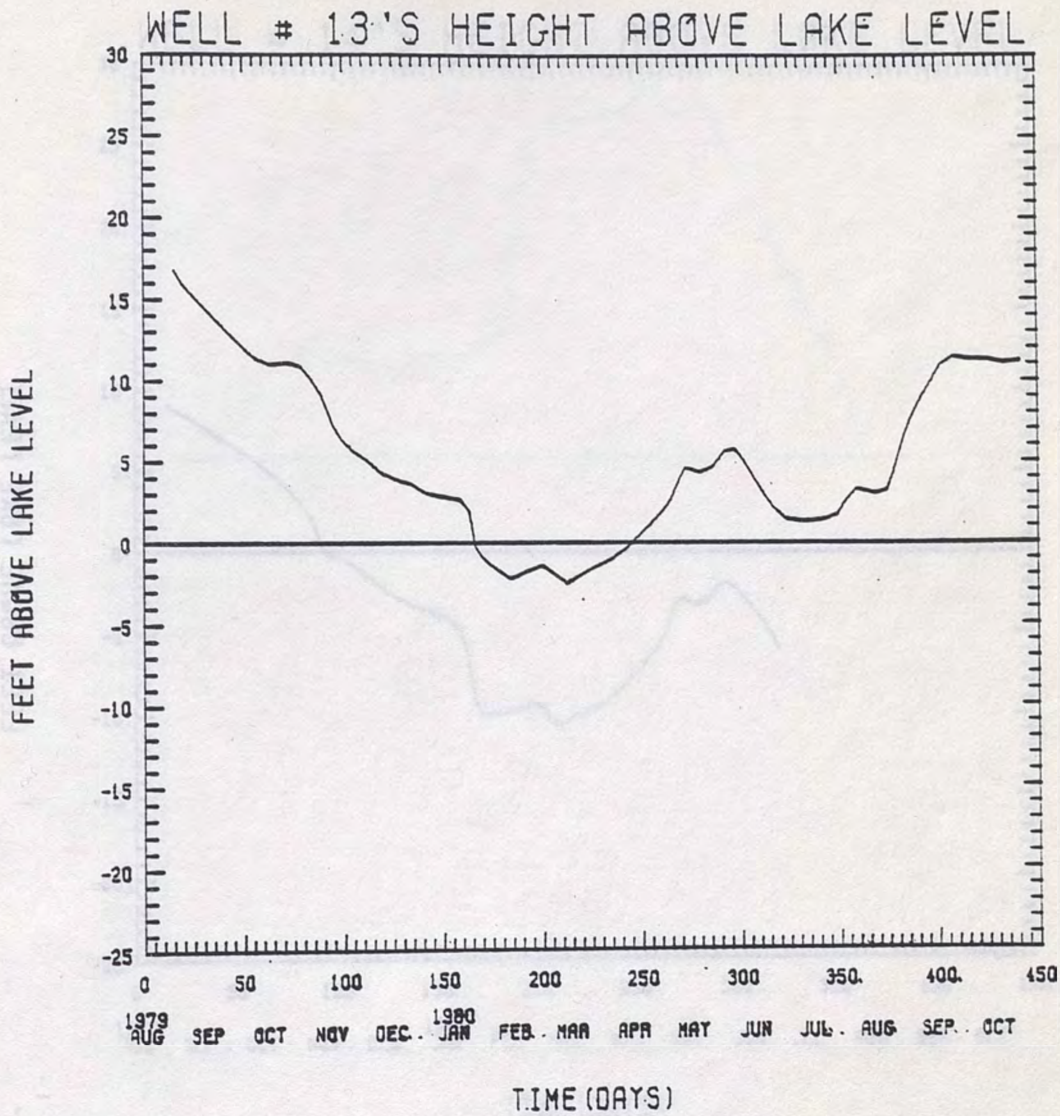


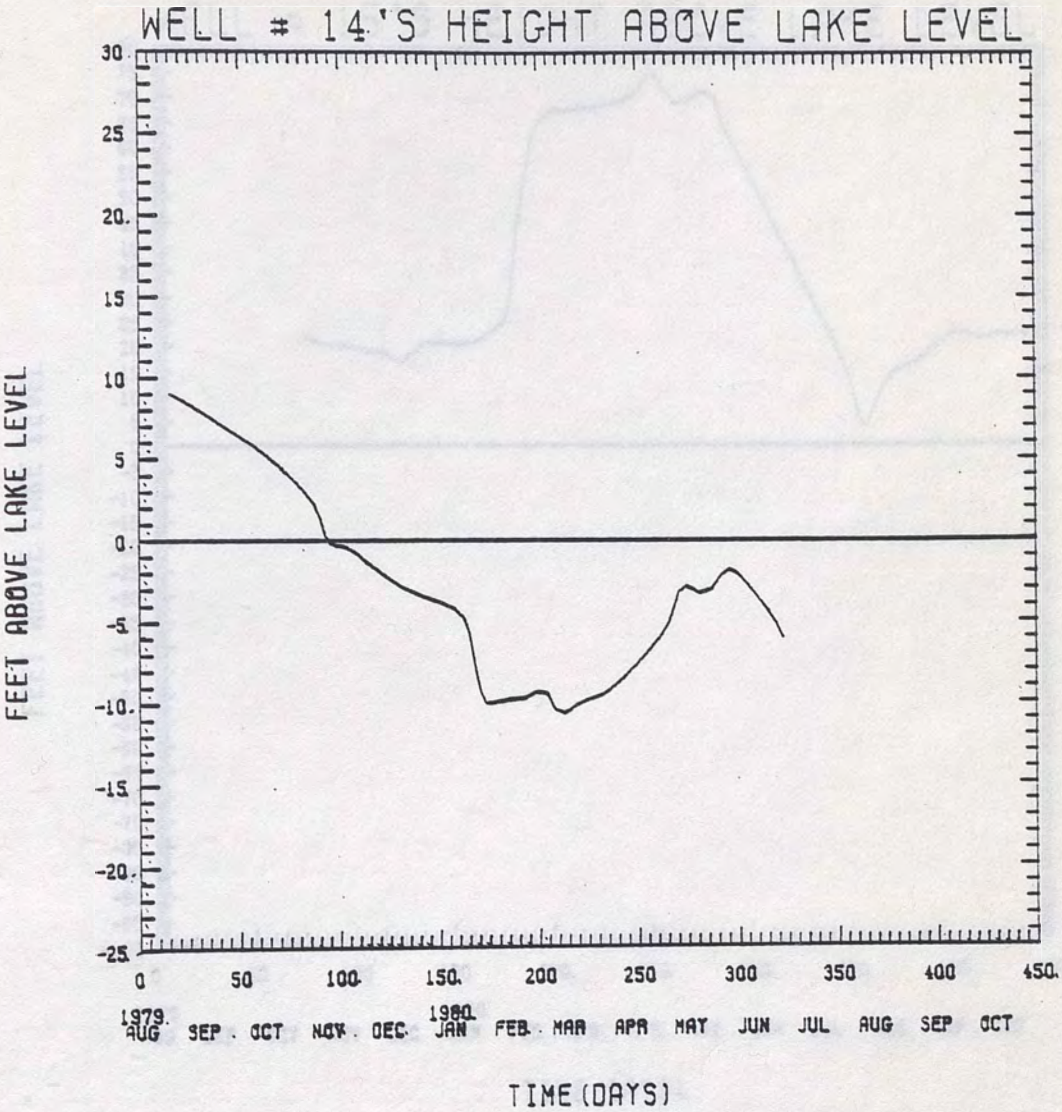
WELL # 9'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
30
25.
20
15
10
5
0.
-5
-10.
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 10'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)



WELL # 12'S HEIGHT ABØVE LAKE LEVEL
FEET ABØVE LAKE LEVEL
30
25.
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
ØCT
NØV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
ØCT
TIME (DAYS)


WELL # 13'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 14'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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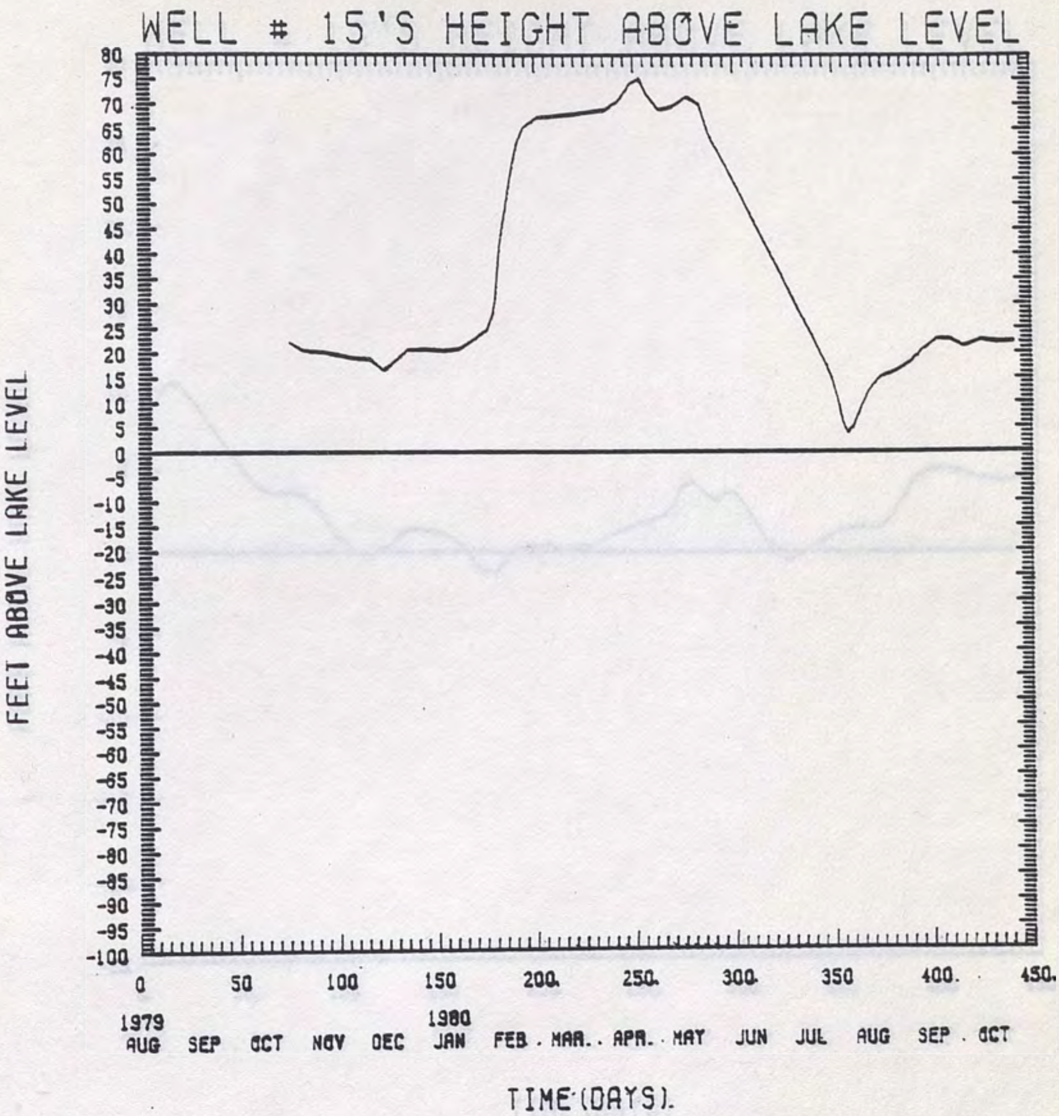
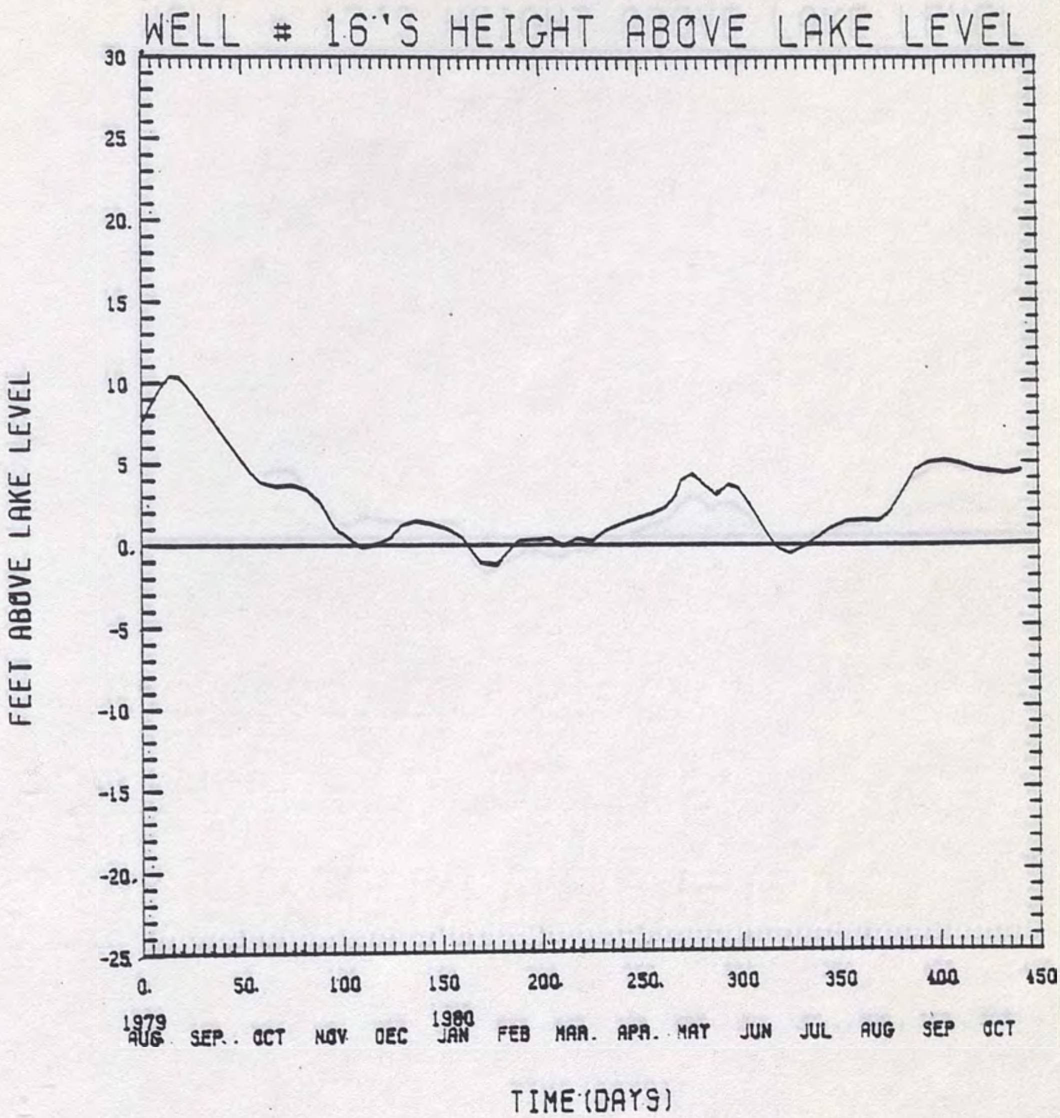
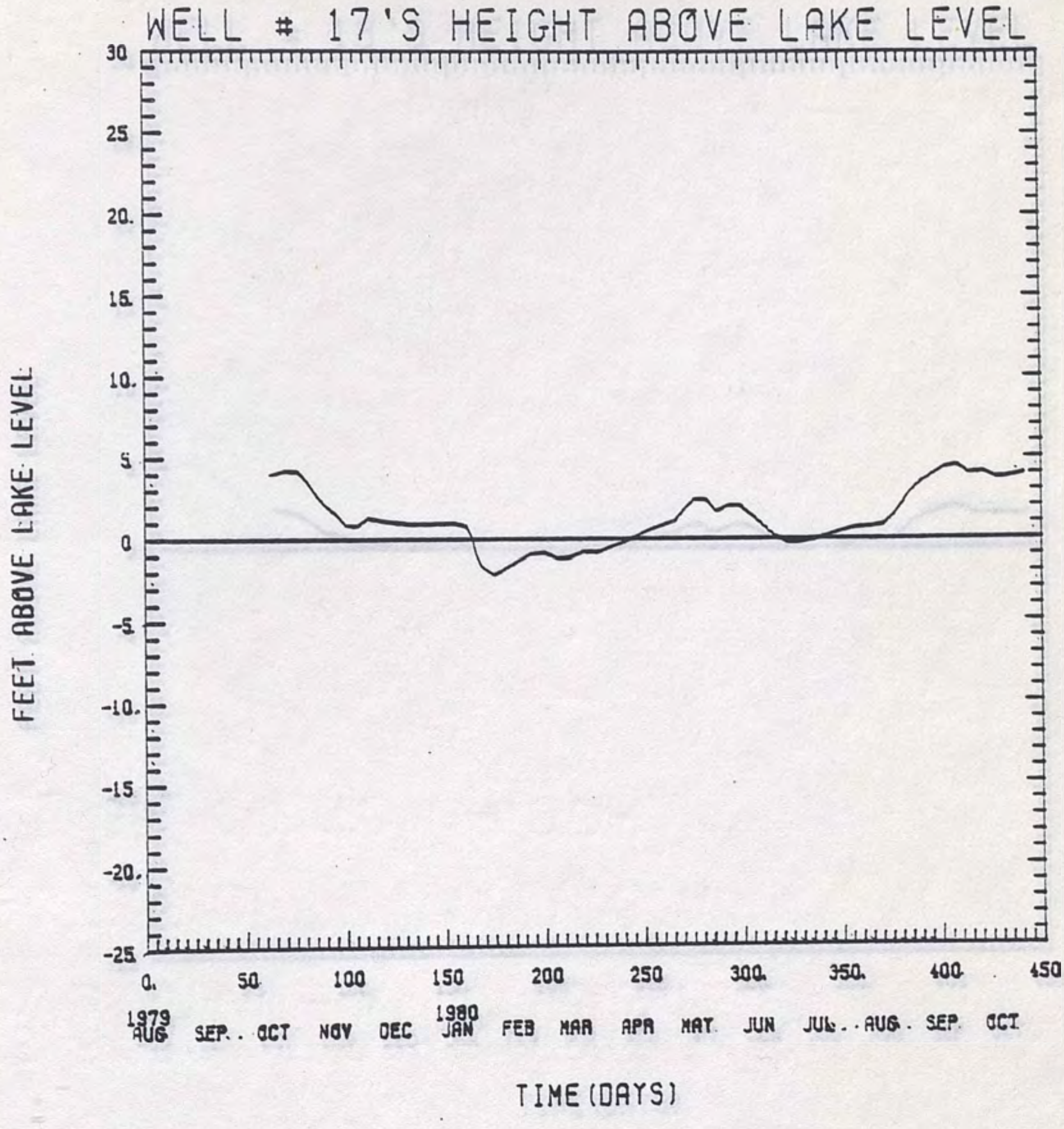
25
20.
15
10
5
0.
-5.
-10.
-15
-20.
-25
0
50
100
150.
200
250
300
350
400
450.
1979.
AUG
SEP
OCT
NOV
DEC
1980.
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)



WELL # 15'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
80
75
70
65
60
55
50
45
40
35
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
-30
-35
-40
-45
-50
-55
-60
-65
-70
-75
-80
-85
-90
-95
-100
0
50
100
150
200
250
300
350
400
450
1979
AUG
SEP
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 16'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
30
25
20.
15
10
5
0.
-5
-10
-15
-20.
-25
0.
50.
100
150
200
250.
300
350
400
450
1979 AUG
SEP.
OCT
NOV
DEC
1980 JAN
FEB
MAR.
APR.
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 17'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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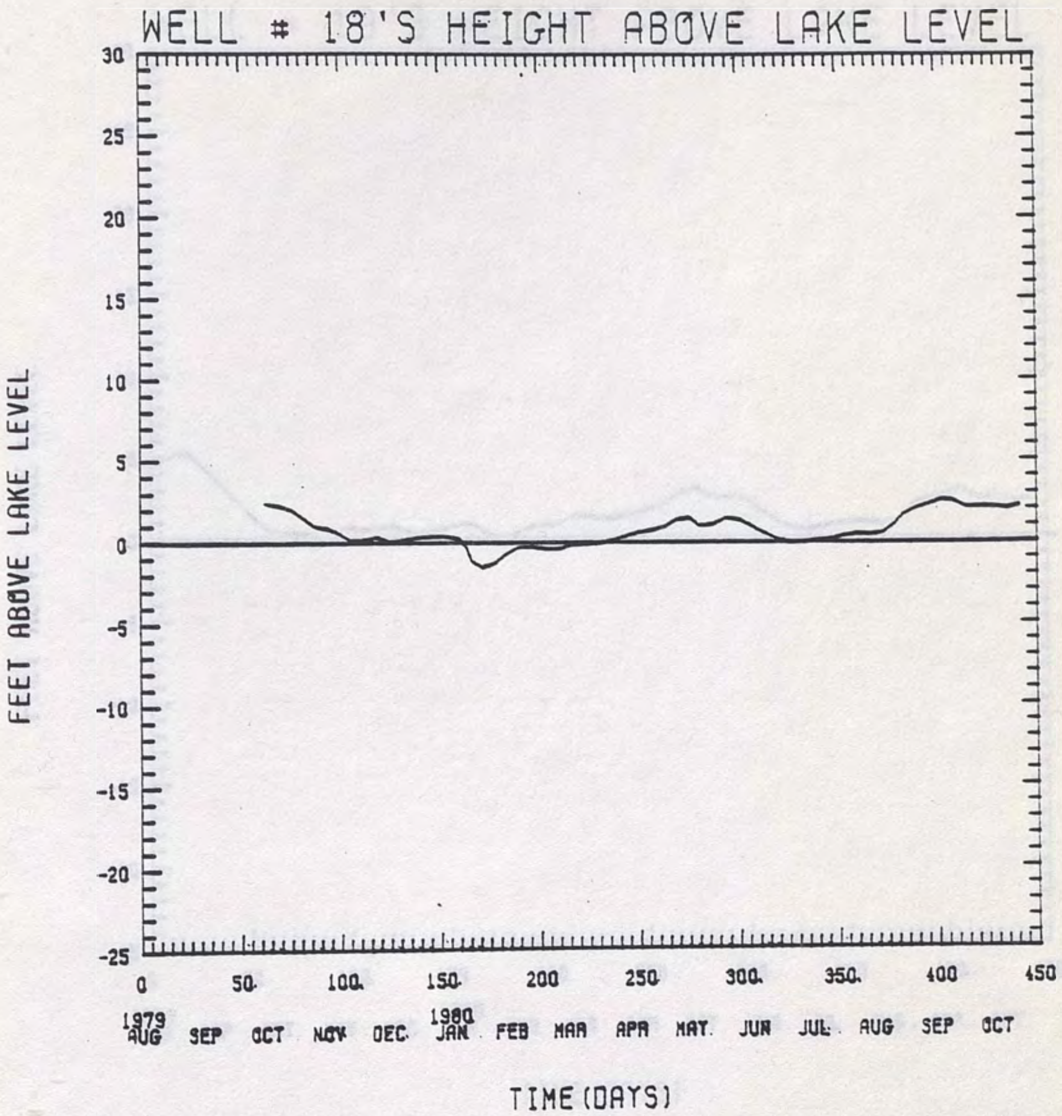
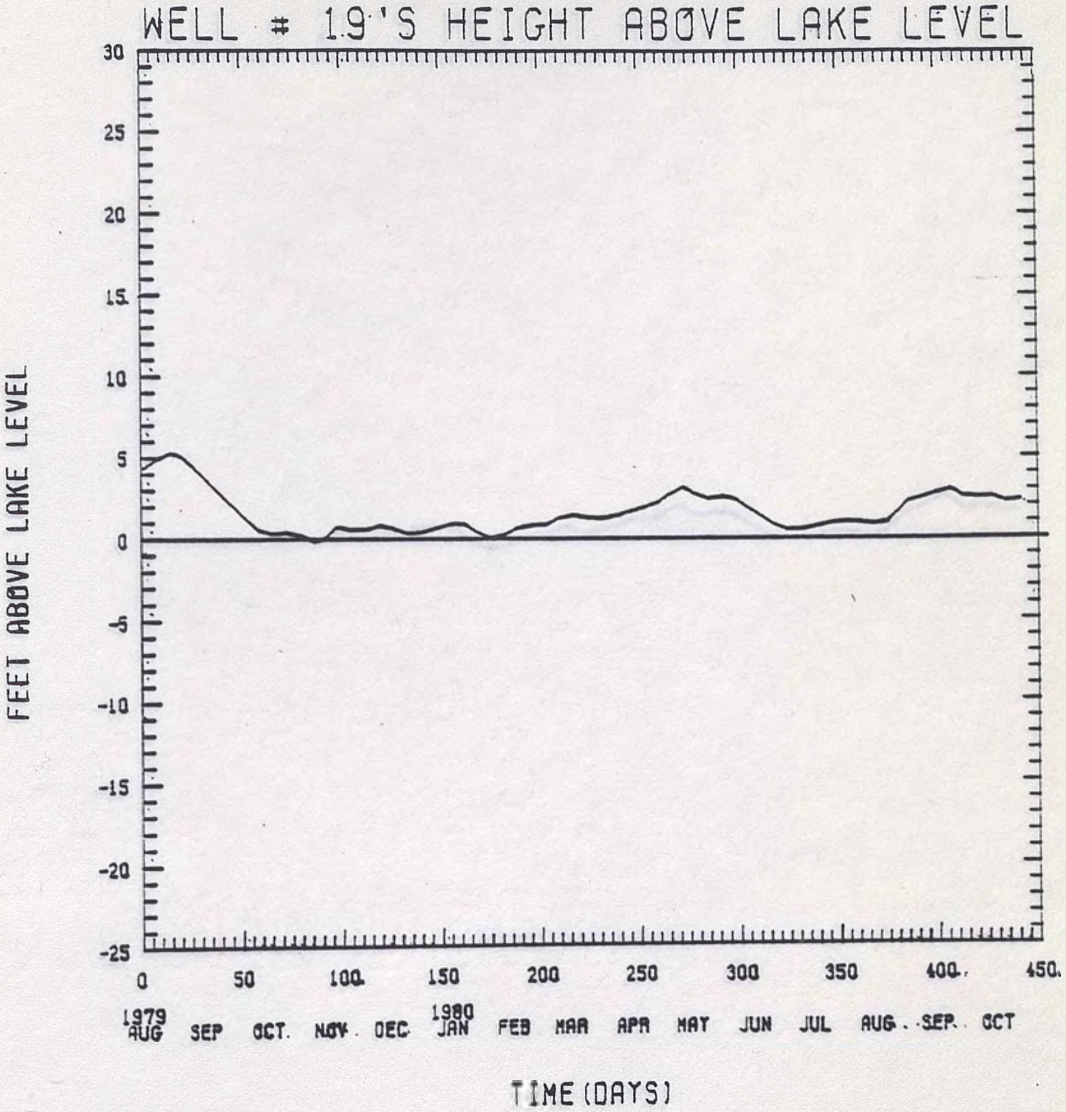
25
20.
15.
10.
5
0
-5.
-10.
-15
-20.
-25
0.
50
100
150
200
250
300.
350.
400
450
1979
AUG
SEP.
OCT
NOV
DEC
1980
JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)



WELL # 18'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
30
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 19'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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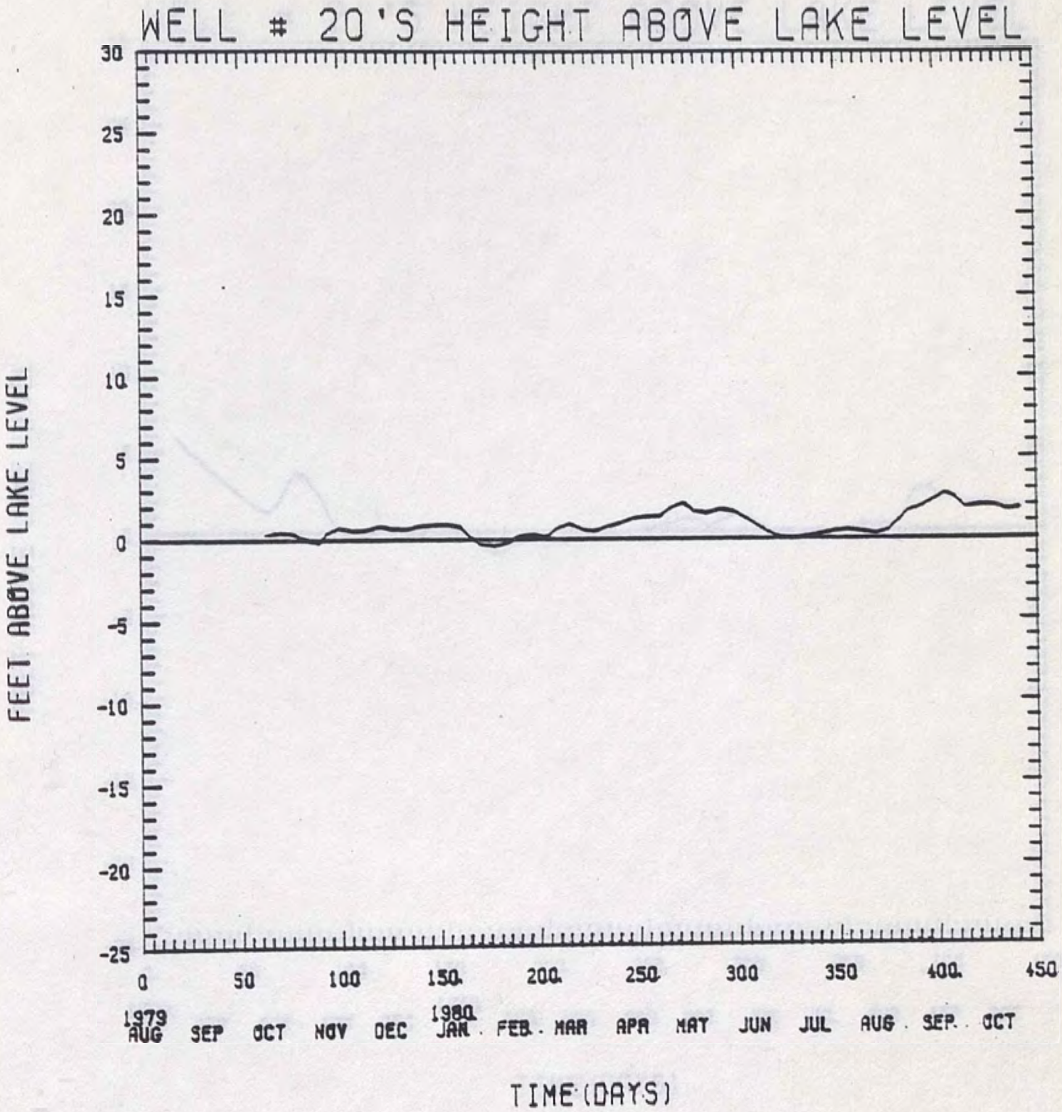
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT.
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 20'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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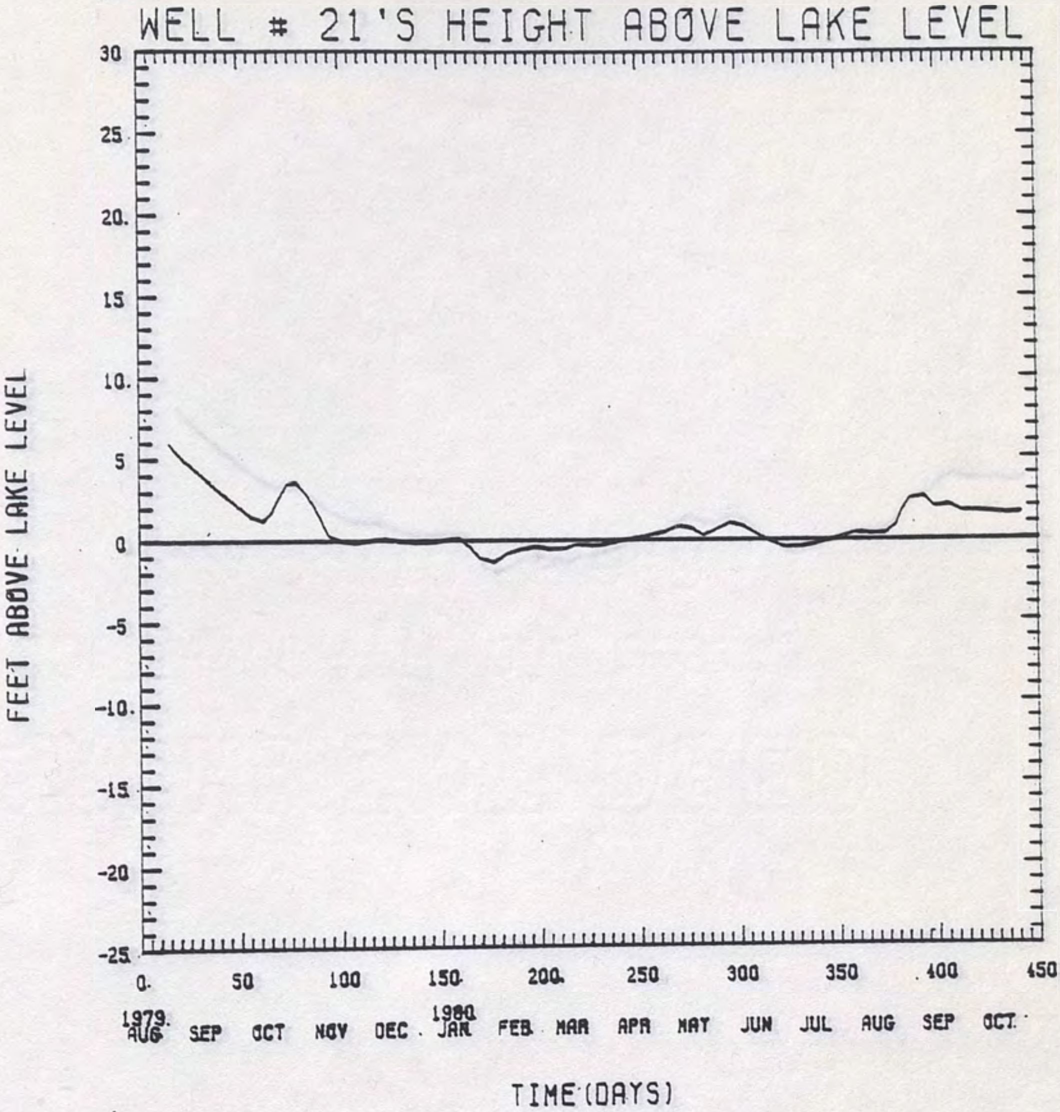
25
20
15
10
5
0
-5
-10
-15
-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
APR
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 21'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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25
20.
15
10.
5
0
-5
-10.
-15
-20
-25
0.
50
100
150
200
250
300
350
400
450
1979 AUG
SEP
OCT
NOV
DEC
1980 JAN
FEB
MAR
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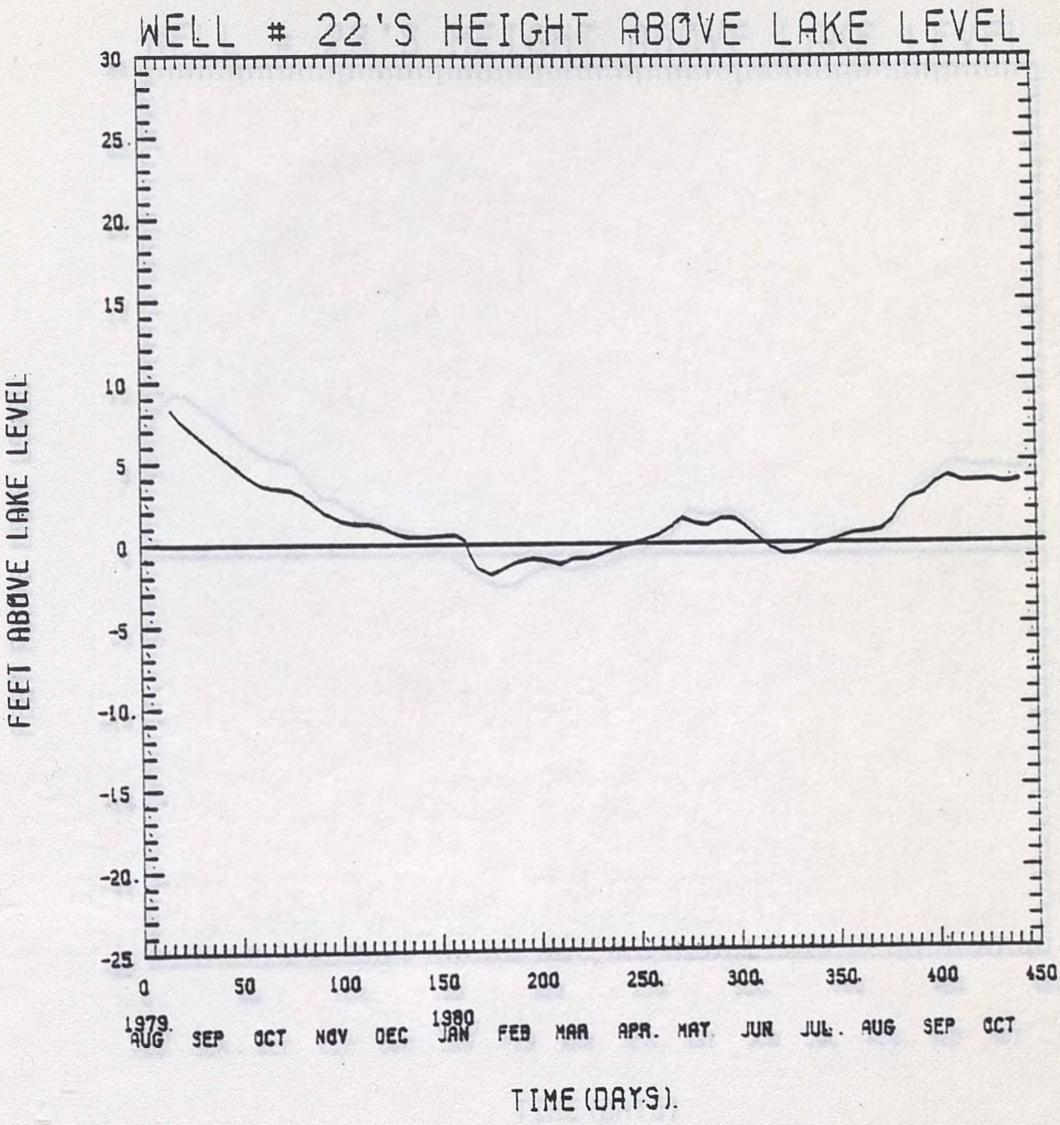
MAY
JUN
JUL
AUG
SEP
OCT.
TIME (DAYS)


WELL # 22'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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25
20
15
10
5
0
-5
-10
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-20
-25
0
50
100
150
200
250
300
350
400
450
1979 AUG
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DEC
1980 JAN
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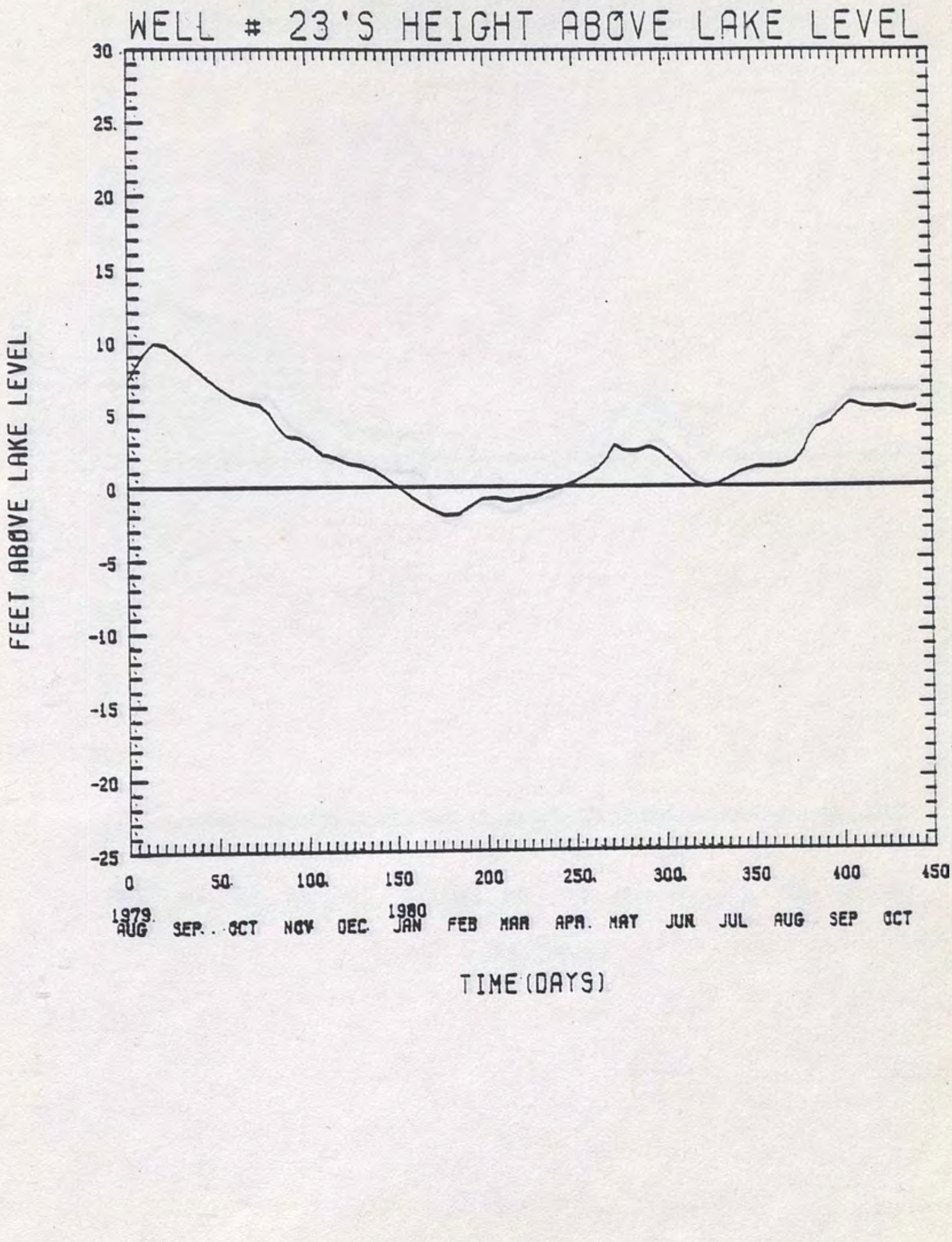
MAY
JUN
JUL.
AUG
SEP
OCT
TIME (DAYS).


WELL # 23'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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25.
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15
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5
0
-5
-10
-15
-20
-25
0
50.
100.
150
200
250.
300.
350
400
450
1979. AUG
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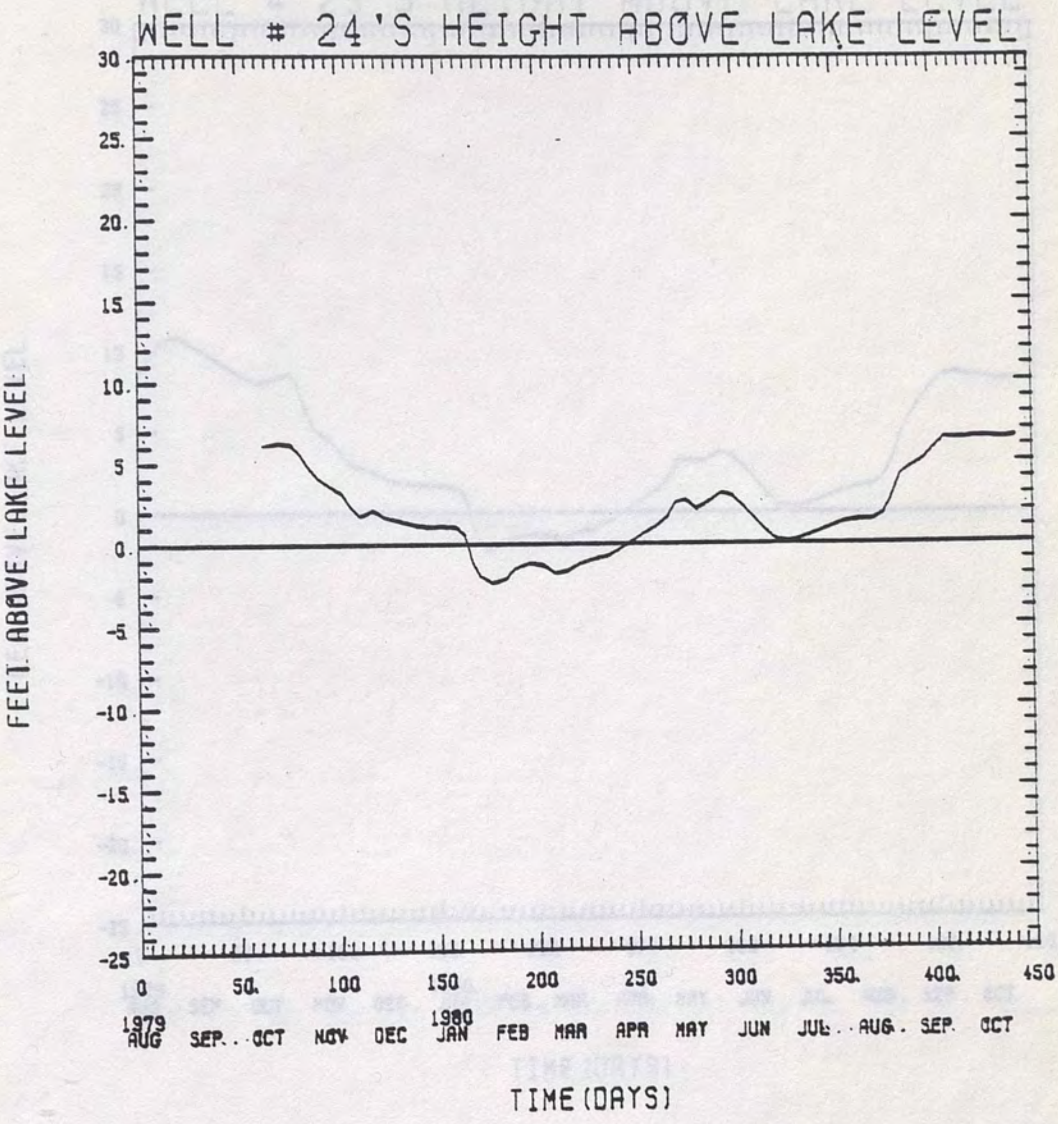
DEC.
1980 JAN
FEB
MAR
APR.
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)



WELL # 24'S HEIGHT ABOVE LAKE LEVEL
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25.
20.
15
10.
5
0.
-5
-10
-15
-20
-25
0
50.
100
150
200
250
300
350.
400.
450
1979 AUG
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1980 JAN
FEB
MAR
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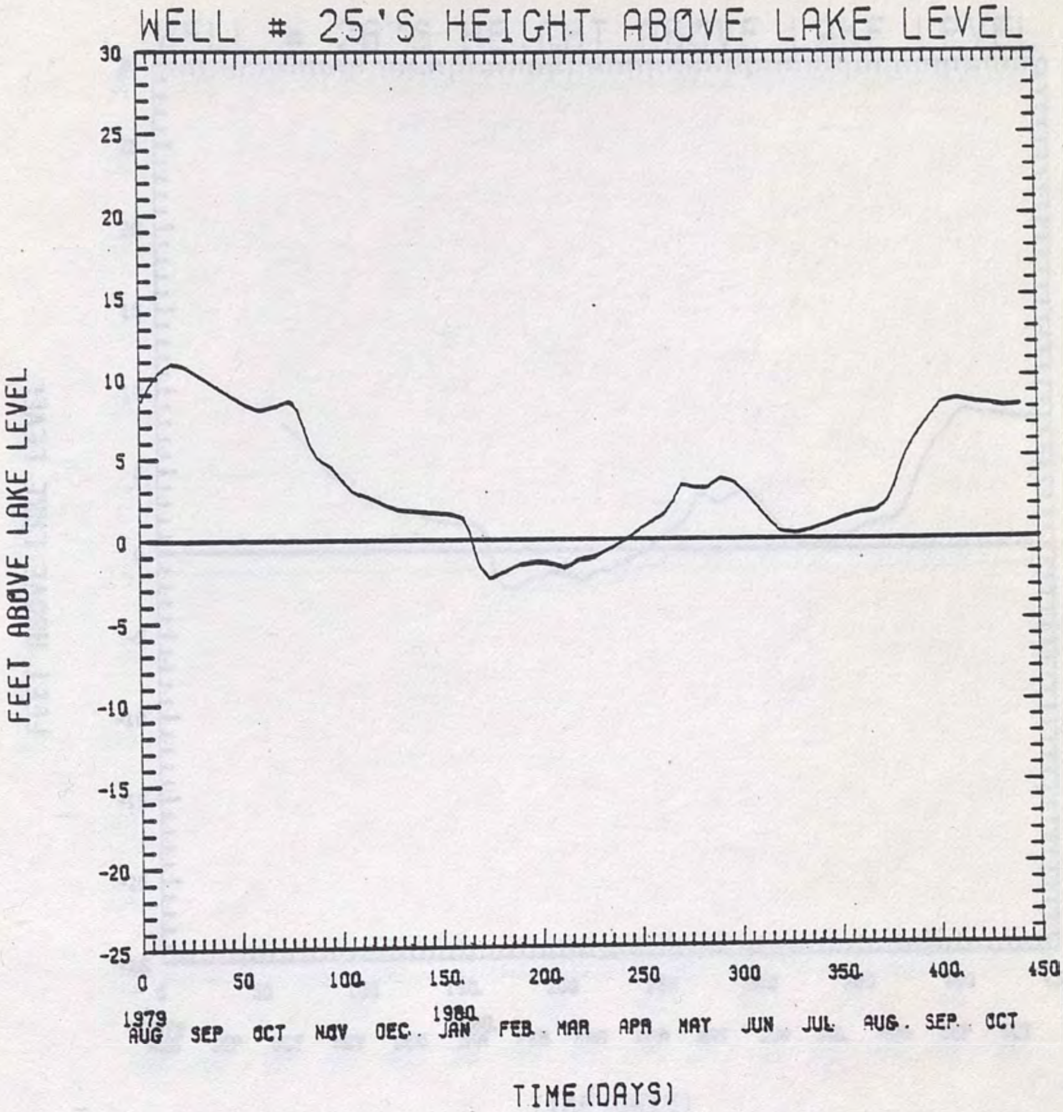
MAY
JUN
JUL
AUG
SEP.
OCT
TIME (DAYS)
FEET ABOVE LAKE LEVEL


WELL # 25'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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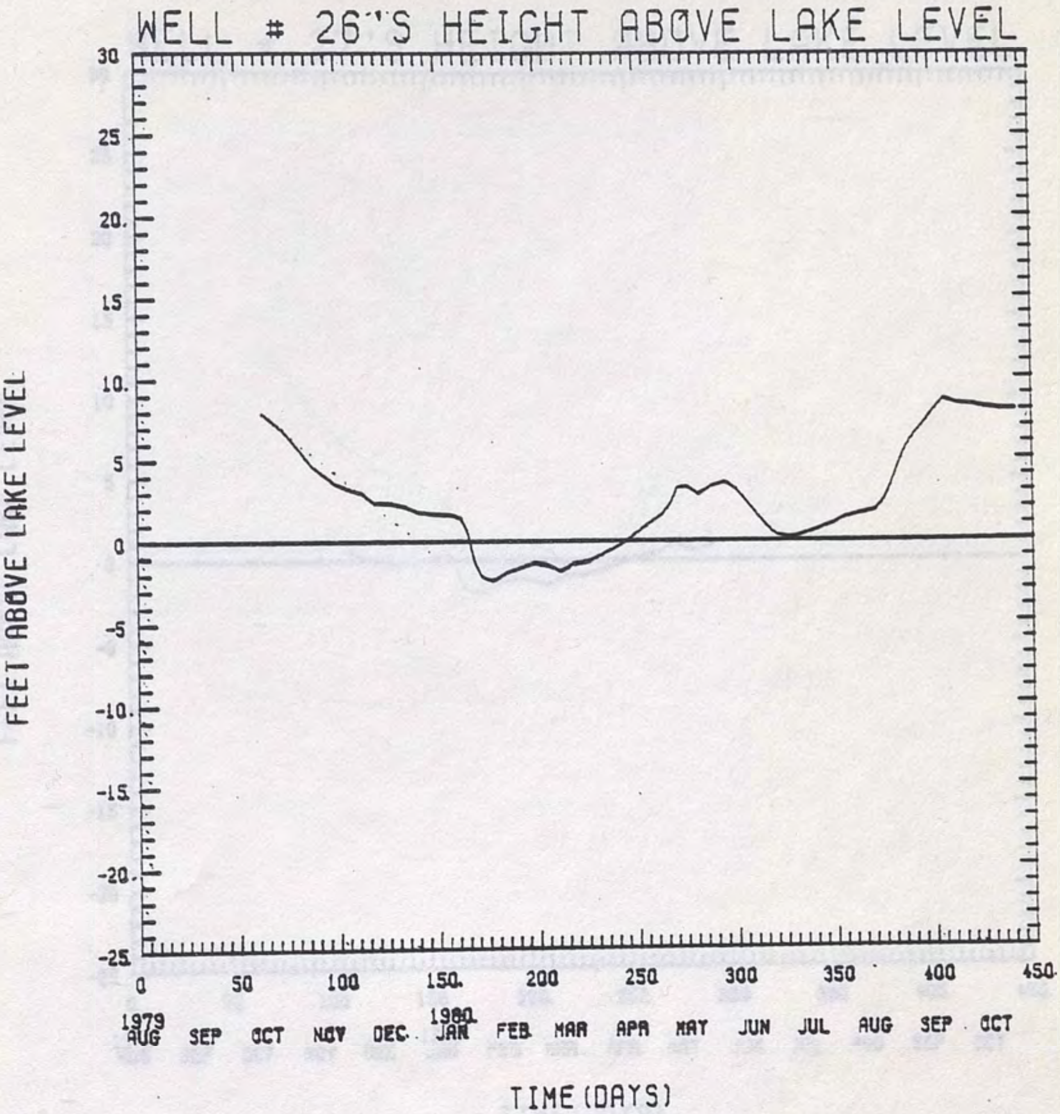
25
20
15
10
5
0
-5
-10
-15
-20
-25
0 50 100 150 200 250 300 350 400 450
1979 AUG SEP OCT NOV DEC 1980 JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)


WELL # 26''S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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20.
15
10.
5
0
-5
-10.
-15
-20.
-25
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50
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150.
200
250
300
350
400
450.
1979 AUG
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DEC
1980 JAN
FEB
MAR
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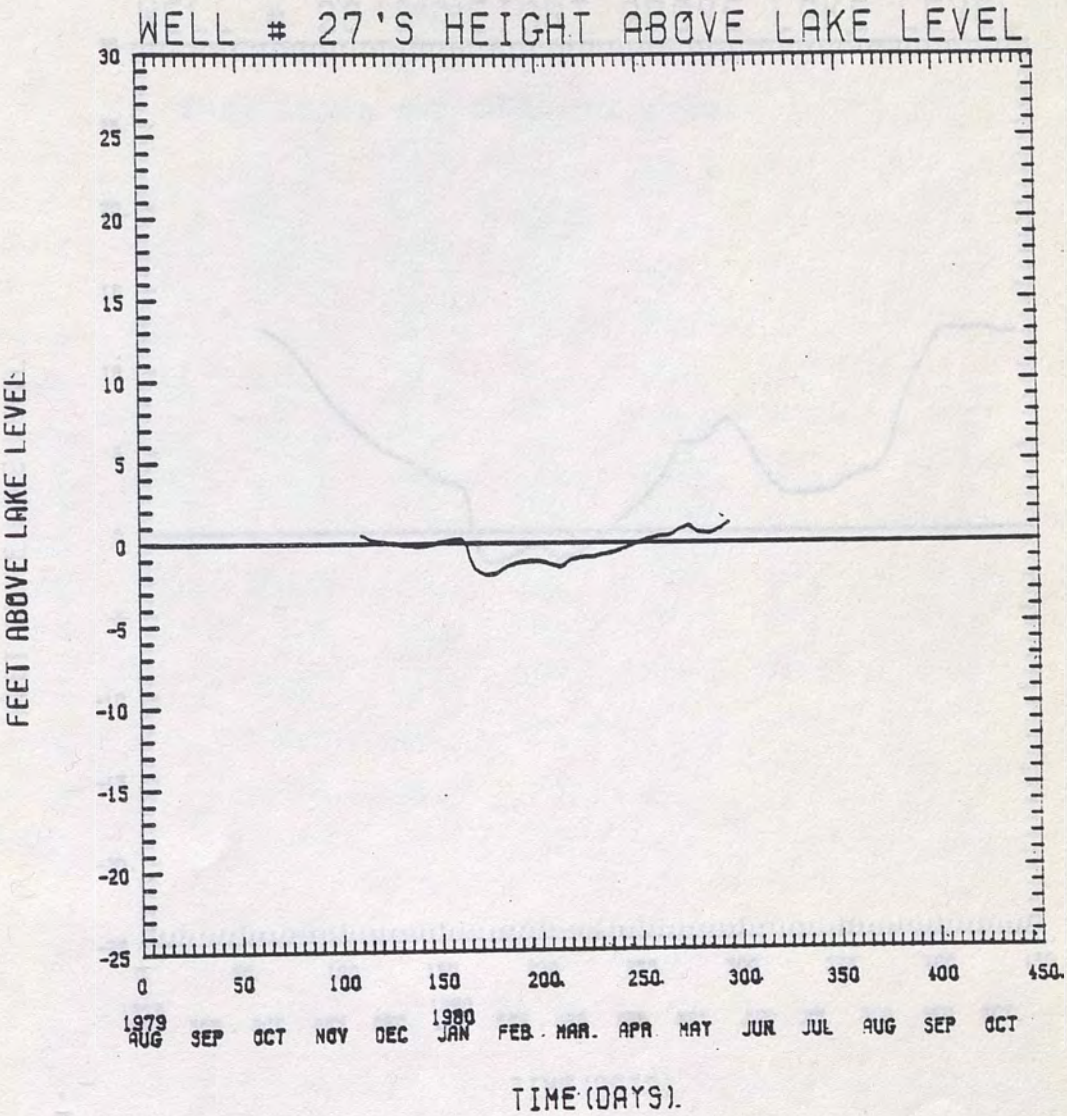
MAY
JUN
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 27'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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25
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15
10
5
0
-5
-10
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-25
0
50
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150
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1979 AUG
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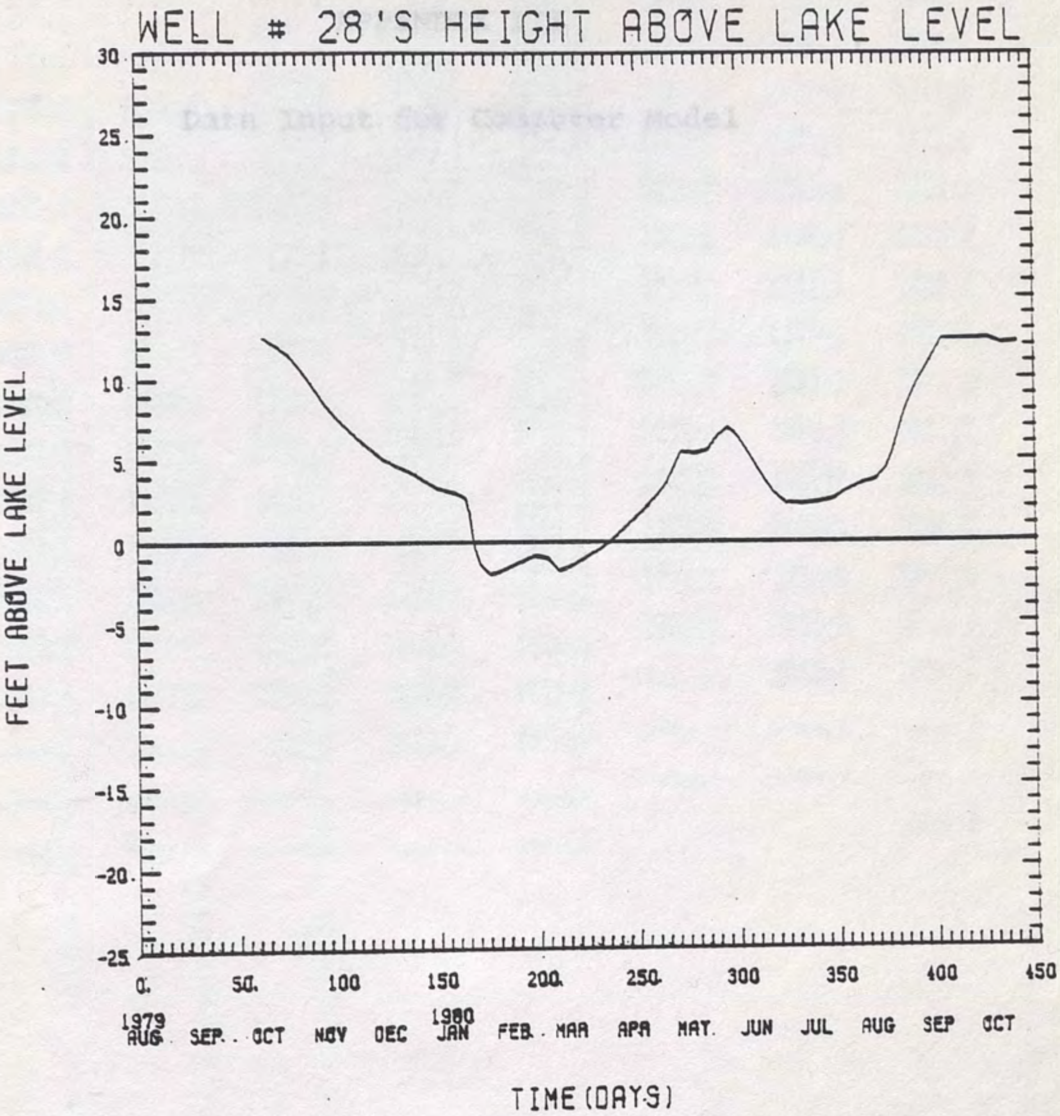
JUL
AUG
SEP
OCT
TIME (DAYS)


WELL # 28'S HEIGHT ABOVE LAKE LEVEL
FEET ABOVE LAKE LEVEL
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25
20
15
10
5
0
-5
-10
-15
-20
-25
0 50 100 150 200 250 300 350 400 450
1979 AUG SEP OCT NOV DEC 1980 JAN FEB MAR APR MAY JUN JUL AUG SEP OCT
TIME (DAYS)

APPENDIX III

Data Input for Computer Model

```
---- FINITE DIFFERENCE PROBLEM WITH CONSTANT HEAD AND RECHARGE WELLS----
WATE                  ADI   CHEC                  NUME HEAD
         20       22                   500
          3      450     .003          .01                                     10
          1        1        1

          1        1


           5002.7   5002.0   5000.1   4999.5   4997.6   4996.5   4996.0
  4995.5   4995.2   4994.9   4994.5   4994.0   4993.6   4993.2   4992.7
  4992.1   4991.5   4990.7   4989.9   4988.8
                    5002.1   5001.2   4999.8   4997.9   4996.6   4996.0
  4995.6   4995.2   4994.9   4994.5   4994.0   4993.5   4993.1   4992.6
  4992.1   4991.4   4990.7   4989.8   4988.8
                    5002.2   5001.4   5000.2   4998.3   4996.7   4996.1
  4995.6   4995.3   4995.0   4994.6   4994.0   4993.5   4993.1   4992.6
  4992.0   4991.4   4990.6   4989.8   4988.8
                    5002.4   5001.6   5000.5   4998.6   4996.8   4996.2
  4995.7   4995.4   4995.0   4994.6   4994.1   4993.5   4993.1   4992.5
  4992.0   4991.3   4990.6   4989.8   4988.8
                    5002.5   5001.9   5000.8   4999.0   4997.1   4996.3
  4995.8   4995.4   4995.1   4994.7   4994.1   4993.5   4993.1   4992.5
  4992.0   4991.3   4990.5   4989.7   4988.8
                             5002.3   5001.2   4999.5   4997.5   4996.5
  4996.0   4995.6   4995.2   4994.7   4994.2   4993.6   4993.1   4992.5
  4991.9   4991.2   4990.5   4989.7   4988.8
                             5002.7   5001.7   5000.1   4998.0   4996.7
  4996.1   4995.7   4995.3   4994.8   4994.2   4993.6   4993.1   4992.5
  4991.9   4991.2   4990.4   4989.7   4988.8
                             5003.1   5002.3   5000.8   4998.8   4997.1
  4996.4   4995.9   4995.5   4995.0   4994.3   4993.7   4993.1   4992.5
  4991.9   4991.1   4990.4   4989.6   4988.8
                             5003.6   5003.1   5001.8   4999.8   4997.9
  4996.9   4996.3   4995.8   4995.2   4994.4   4993.7   4993.2   4992.5
  4991.9   4991.1   4990.4   4989.6   4988.8
                             5004.1   5004.3   5003.1   5001.3   4999.4
  4997.8   4996.8   4996.2   4995.5   4994.7   4993.9   4993.3   4992.6
  4991.9   4991.1   4990.4   4989.6   4988.8
                                      5006.7   5005.0   5003.3   5001.6
  4999.9   4998.2   4997.1   4996.1   4995.1   4994.2   4993.5   4992.7
  4992.0   4991.2   4990.4   4989.6   4988.8
                                      5008.1   5006.4   5004.9   5003.4
  5001.9   5000.3   4998.6   4996.8   4995.5   4994.5   4993.7   4992.8
  4992.0   4991.1   4990.4   4989.7   4988.8
                                      5008.8   5007.6   5006.4   5005.2
  5003.9   5002.4   5000.7   4998.4   4996.3   4994.8   4993.9   4993.0
  4992.1   4991.2   4990.4   4989.6   4988.8
                                               5008.7   5007.8   5006.9
  5005.7   5004.4   5002.7   5000.5   4997.6   4995.2   4994.2   4993.1
  4992.1   4991.2   4990.4   4989.6   4988.8
                                                        5008.7   5008.1
  5007.1   5006.0   5004.0   5002.0   4999.2   4996.2   4994.5   4993.2
  4992.1   4991.2   4990.4   4989.6   4988.8
                    5004.9   5003.8   5001.3   4998.7   4996.3   4993.7
  4992.1   4991.1   4990.4   4989.6   4988.8
                                               4999.7   4998.2   4994.9
  4992.4   4991.0   4990.3   4989.0   4988.8
                                                                 4995.3
  4992.8   4990.8   4990.4   4989.5   4988.8

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        70   70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
        70   70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
        70   70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
             70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
             70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
             70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
             70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50   50
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                  70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50
                  70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50
                  70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50   50
                       70   70   70   70   70   50   50   50   50   50   50   50   50   50   50   50
                            70   70   70   70   70   50   50   50   50   50   50   50   50   50   50
                                               70   70   70   70   50   50   50   50   50   50   50
                                                              70   70   50   50   50   50   50   50
                                                                        70   70   50   50   50   50
```

```
        1        1
49704970497049704910482047604700466046204575452044854515455545654575461546604705
    4970497049704960487047904735470046604600455045154480451045454555456546104655
    4970497049704975488048154770472046804640460545604530446545004535454545604605
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        497049704970497049704950491048804855482047804730470546754625456545204495
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                                                497049704970493548104760470046854715
                                                        497049704010480047404715
```

```
  .15
   .1          1
 2500       2500       2500       2500       2500       2500       2000       1500
 1500       1000       1500       1500       2500       1000       2000       1500
 2000       2000       2000       2000       2000       1500
   .1          1
 3500       1500       2000       1500       1000       1500       1500       1000
 2000       1500       2500       2500       1500       2500       1000       1500
 2500       2500       2500       2500
    1          0         24        168        168        1.0         24
    2          2        .01
    3          3        .01
    4          3        .01
    5          3        .01
    6          3        .03
    7          4        .03
    8          4        .05
    9          4        .05
   10          4        .03
   11          4        .03
   12          5        .07
   13          5        .07
   14          5        .07
   15          6        .07
   16          7        .07
   16          8        .07
   16          9        .07
   15         10        .07
   17         11        .07
   17         12        .07
   17         13        .07
   18         14        .07
   18         15        .07
   19         16        .07
    2          1         24        140        140        1.0         24
    2          2        .04
    3          3        .04
    4          3        .04
    5          3        .04
    6          3        .07
    7          4        .07
    8          4        .07
    9          4        .07
   10          4        .07
   11          4        .07
   12          5        .07
   13          5        .07
   14          5        .07
   15          6        .07
   16          7        .07
   16          8        .07
   16          9        .07
   16         10        .07
   17         11        .07
   17         12        .07
   17         13        .07
   18         14        .07
   18         15        .07
   19         16        .07
```

3	2	24	119	119	1.0	24
2	2	.01				
3	3	.01				
4	3	.01				
5	3	.04				
6	3	.03				
7	4	.06				
8	4	.09				
9	4	.09				
10	4	.06				
11	4	.06				
12	5	.03				
13	5	.03				
14	5	.02				
15	6	.02				
16	7	.01				
16	8	.01				
16	9	.01				
16	10	.01				
17	11	.01				
17	12	.01				
17	13	.01				
18	14	.01				
18	15	.01				
19	16	.01				